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Editor-in-Chief

Boris S. Sokolov

Academician, Russian Academy of Sciences,
Counsellor of the Presidium of the Russian Academy of Sciences

Address for correspondence: Staromonetnyi per. 22, Moscow 109180 Russia
Phones: 7-(095) 231-2164; Fax: 7-(095) 233-5590.

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Changes in the Taxonomic and Ecologic Composition of Benthic Shelf Assemblages at the Cretaceous–Paleogene Boundary

L. A. Nevesskaya

Paleontological Institute, Russian Academy of Sciences, Profsoyuznaya ul. 123, Moscow, 117647 Russia

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Abstract – Changes in the taxonomic and ecologic composition of benthic shelf assemblages at the Cretaceous–Paleogene boundary and the conceivable reasons for these changes are discussed. Analysis of the available data shows that, although mass extinction and general crisis of the biota at this boundary were among the most considerable Phanerozoic events, they were asynchronous and variable in scale for different benthic groups. The ethological and trophic compositions were stable in most shelf zones; only reef communities composed of sessile sestonophages disappeared.

Key words: *benthic communities, taxonomic and ecologic composition, Cretaceous–Paleogene boundary, crisis, extinction.*

The benthos of the Cretaceous shelf seas was dominated by bivalve mollusks usually in association with foraminifers, sponges, corals, hydroids, brachiopods, bryozoans, gastropods, scaphopods, ostracodes, cirripeds, malacostracans, and other crustaceous as well as echinoids, crinoids, ophiuroids, asteroids, holothuroids, and worms (namely, polychaets and priapulids). Stromatoporoids, polyplacophorans, and some other groups occurred more rarely. The bottom vegetation consisted of diverse green, blue-green, and red algae.

The nektonic and nektobenthic groups, which partly served as benthos feeders, included cephalopods, fishes, and reptiles (Bambach, 1983; Thayer, 1983; Shimanskii, 1987; and others).

Plankton, the nourishment for many benthic forms, was abundant and diverse. Dinoflagellates and coccolithophores dominated among the phytoplankton. In the Middle Cretaceous, diatoms and silicoflagellates appeared, and they flourished through the Late Cretaceous, whereas other groups were of secondary importance. Radiolarians dominated among the zooplankton, whereas planktonic foraminifers prevailed in the second half of the Cretaceous. Tintinnids also occurred and were numerous at the beginning of this period, although their abundance later decreased (Downie, 1967; Tappan, 1968; Tappan and Loeblich, 1973; Fischer and Arthur, 1977; *Plankton i Organicheskii ...*, 1979; Kemper, 1982; Benda, 1982; Knoll, 1989).

Some researchers consider the second half of the Early Cretaceous (Barremian–Aptian) to be the time during which the modern ocean appeared, giving rise to pelagic fauna and flora of the modern type, namely, the mass development of planktonic foraminifers and diatoms as well as other pelagic forms of phyto- and zooplankton (Taylor, 1980; Kemper, 1982; Benda, 1982).

Several episodes of extinctions and changes in the organic life occurred during the Cretaceous: in the Aptian, the Cenomanian–Turonian boundary period, and the Maestrichtian.

However, a major crisis of the biota, including benthos, occurred during the Maestrichtian. Planktonic forms experienced the most intense extinction. Among the calcareous nannoplankton, about 85% of species, 70% of genera, and 30% of families became extinct. The mass extinction involved silicoflagellates, dinoflagellates, coccolithophores, planktonic foraminifers, and radiolarians, whereas diatoms and prasinophytes survived this stage without substantial changes. A rapid extinction was also recorded for plankton-dependent nektonic forms, i.e., for ammonoids, belemnoids, marine reptiles, and bony fishes.

The following benthic organisms suffered the most: bivalve mollusks, among which Hippuritida and many other families disappeared; brachiopods; sponges; stromatoporoids; bryozoans; and echinoids. At the same time, gastropods, scaphopods, crinoids, ostracodes, and some groups of bivalve mollusks did not change significantly at the Cretaceous–Paleogene boundary (Bramlette and Martini, 1964; Tappan, 1968, 1982; Tappan and Loeblich, 1972; Newell, 1972; Moskvin *et al.*, 1972; Naidin, 1976; Saks, 1976; Russell, 1977; Fischer and Arthur, 1977; *Cretaceous–Tertiary ...*, 1979; Bujak and Williams, 1979; Moskvin, 1979, 1989; *Razvitie i Smena Bespozvonochnykh ...*, 1980; *Razvitie i Smena Mollyuskov ...*, 1981; *Razvitie i Smena Prosteishikh ...*, 1981; Emiliani *et al.*, 1981; Shimanskii and Solov'ev, 1982; *Cretaceous–Tertiary Extinctions ...*, 1982; Perch-Nielsen *et al.*, 1982; Sepkoski, 1982; Thierstein, 1982; Ward, 1983; Smit and Romein, 1985; McKinney, 1985; Krasilov, 1985; Sepkoski and Raup, 1986; Masse and

Philip, 1986; Perch-Nielsen, 1986; Alekseev, 1986; Lipps, 1986; Jones and Nicol, 1986; Sheehan and Hansen, 1986; Shimanskii, 1987; Hansen *et al.*, 1987; Keller, 1988a; House, 1989; Knoll, 1989; Bryan and Jones, 1989; Hallam and Perch-Nielsen, 1990; Olssen and Liu, 1993).

It should be noted that shallow-water sestonophages, which were dependent on the organic matter input from the euphotic zone, became extinct to a greater degree than other benthic organisms. Detritophages and carnivores suffered to a lesser extent. A similar situation was suggested for deep-water benthos, i.e., for ostracodes and foraminifers, which had adapted to scanty nutrition resources (Arthur *et al.*, 1987; Hansen *et al.*, 1987; Rhodes and Thayer, 1991).

Altogether, 60 - 75% of species, more than 35% of genera, and about 20% of families had become extinct at the end of the Cretaceous (Ager, 1981; Sepkoski, 1986, 1988, 1989). Along with them, about 50% of benthic, 35% of planktonic, and 80% of nektonic marine genera disappeared (*Cretaceous–Tertiary Extinction ...*, 1982).

However, certain groups of biota experienced degradation before the end of the Maestrichtian; thus, the extinction events were not synchronous. For example, the degradation of ammonoids began in the Coniacian, and only certain long-existing forms occurred during the Maestrichtian. Similar data are available for inoceramids (bivalve mollusks) as well as for dinosaurs, mammals, and land vegetation (Clemens *et al.*, 1981; Thierstein, 1982; Shimanskii and Solov'ev, 1982; Ward, 1983; Kauffman, 1984; Smit and Romein, 1985; Buffetaut, 1985; Ward *et al.*, 1986; Jaeger, 1986; Wiedmann, 1986; Roth, 1987; Officer *et al.*, 1987; Hansen *et al.*, 1987; Jones *et al.*, 1987; Ward and MacLeod, 1988; Kennedy, 1989; Hoffman, 1989; Holland, 1989; Keller, 1989; Kennedy, 1989; Knoll, 1989; Rhodes and Thayer, 1991; MacLeod and Keller, 1991).

The intense extinction and change in the biota at the Cretaceous–Paleogene boundary most likely resulted from a decline in primary productivity, which caused an abrupt decrease in nutritient resources (Bramlette, 1965; Tappan, 1968; Zachos *et al.*, 1988, 1989; Rhodes and Thayer, 1991; and others). However, there are still no generally accepted explanations for this decline in productivity. Two main groups of hypotheses, extraterrestrial and terrestrial, have been advanced.

Supporters of the extraterrestrial hypothesis, i.e., of the impact hypothesis involving the fall of a bolide or the passage of the Earth through comet clusters ("comet rain"), suggest that these phenomena resulted in the formation of a large dust cloud, which darkened the sun and prevented photosynthesis for several years. The latter, in turn, caused the mass extinction of phytoplankton and then the extinction of phytoplankton feeders. The effect of darkening was reinforced by acidic rains (Alvarez *et al.*, 1980, 1984; Hsü, 1980, 1982, 1989; Smit and Hertogen, 1980; Ganapathy, 1980; Hsü *et al.*, 1982; Pollack *et al.*, 1983; Wolbach *et al.*, 1985, 1988; Alekseev, 1986; Hut *et al.*, 1987; Prinn and Fegley, 1987; Van den Berch, 1989; Smit *et al.*, 1988; Alekseev *et al.*, 1990; Kaiho, 1992; a others). An experimental test of this hypothesis on modern phytoplankton confirmed the possible effect of darkening and acidic rains, although it was not as pronounced as supporters of the impact event had postulated (Griffis and Chapman, 1988, 1990). Therefore, sufficient evidence in favor of the impact hypothesis of extinction is lacking. At the same time, the selective and gradual character of the extinction contradicts this hypothesis. Some faunal groups became completely extinct, others became extinct only partially, and some groups did not suffer at all. In addition, these events, as noted above, were not synchronous (Russell, 1977; McLean, 1985; Smit and Romein, 1985; Hansen *et al.*, 1986; Hallam, 1987; Officer *et al.*, 1987; Johansen, 1989; Keller, 1989; and others). As for geochemical and physical evidence, iridium peaks were recorded both above and below the Cretaceous–Paleogene boundary and may have resulted from reinforced volcanic activity, relevant microspherùls could conceivably be of organic or volcanic origin, and quartz textures that are considered to be of impact origin may have been produced by metamorphism (Officer and Drake, 1983, 1985, 1989; Officer *et al.*, 1987; Sahni, 1988; Hoffman, 1989; Officer, 1990; Zakharov and Lapukhov, 1991).

Newell's hypothesis was the first to suggest terrestrial reasons for the extinction at the end of the Cretaceous. It assumes a disruption in oceanic circulation; rapid burial of nutrients; and the occurrence of a stagnation, i.e., oxygen deficiency in the water (Newell, 1965). This hypothesis has been supported by other researchers (Zachos and Arthur, 1986; Brinkhuis and Zachariasse, 1988; Zachos *et al.*, 1989) who recorded a decline in the primary productivity of surface waters, which was most likely caused by the termination of intense water circulation.

According to some investigators, the extinction may also have resulted from the following: (a) A deep global regression occurred at the end of the Cretaceous, which caused an expansion of sea areas, transformed the bottom topography and character of sedimentation, and provoked other environmental changes as well as the extinction of benthic organisms (Wiedmann, 1969; Cooper, 1977; Hallam, 1984; Ginsburg, 1986; Jaeger, 1986; House, 1989). (b) Volcanic activity intensified (Officer and Drake, 1983, 1985; Campsie *et al.*, 1984; McLean, 1985, 1988; Officer *et al.*, 1987; Courtillot and Cisowski, 1987; Erickson III and Dickson, 1987; Crocket *et al.*, 1988; Sahni, 1988) and was accompanied by phenomena similar to those of the impact event, namely, acidic rains, acidification of oceanic surface waters, and dispersal of toxic trace elements. This resulted in the destruction of the calcareous plankton, while the eruption products, dust and aerosols, may have disturbed the ozone screen, darkening the atmosphere and causing cooling or resulting in

a greenhouse effect due to the increased content of carbon dioxide in the atmosphere. (c) A combination of regression and volcanism occurred (Ekdale and Bromley, 1984; Hallam, 1987, 1989; Officer *et al.*, 1987; Officer and Drake, 1989). (d) The upper water layer in oceans was freshened due to water input from the Arctic basin, and the oxygen content decreased below the upper layers. This resulted in the extinction of planktonic forms due to the freshening, the extinction of benthic fauna due to the oxygen deficiency, and the extinction of land organisms due to temperature decline and the lowering of precipitation (Gartner and Keany, 1978; Thierstein and Berger, 1978; Gartner and McGuirk, 1979; Thierstein, 1982). (e) Cooling (Stanley, 1984) was recorded by an oxygen-isotope analysis of planktonic and benthic foraminifers, which may have led to an intensification of ocean circulation and a temperature decline and oxygen deficiency due to the mixing of shelf and oxygen-depleted deep waters (Wilde and Berry, 1986). On land, the results would have been less intense weathering and a decrease in the amount of nutritients being input into the seas (Bramlette, 1965). (f) Oceanic water heterogeneity was reduced due to a decrease in the thermal gradient (Lipps, 1986). (g) Biotic reasons, namely, an abundance of deciduous vegetation increased the soil cover and decreased the amount of nutrients entering seas and oceans, which would have resulted in the extinction of phytoplankton followed by the extinction of plankton consumers (Bramlette and Martini, 1964; Bramlette, 1965; Tappan, 1982; Sheehan and Hansen, 1986; Arthur *et al.*, 1987; Gallagher, 1991). From another point of view, the depletion of oceanic nutrient resources could also result from extremely high phytoplankton productivity and the development of "red" tides (Naidin, 1986).

Less common hypotheses suggest that the extinction of plankton and other organisms resulted from the following: the elevation of the carbonate compensation level to the surface (Worsley, 1971); climate warming (Emiliani *et al.*, 1981; Rampino and Volk, 1988); geomagnetic reversals that caused genetic instability, the appearance of mutants, and a decline in the adaptability of some groups of marine and land organisms (Tsakas and David, 1987); and, finally, a rise in gravity, which was caused by the crossing of a galaxian gravitational wave front by the Solar System (Belkin, 1989).

Following the crisis, the reestablishment of species and genera diversity began during the second half of the Paleocene. However, it was not until the middle Eocene that the majority of benthic groups, namely, foraminifers, bivalve mollusks, gastropods, ostracodes, and echinoids, reached or exceeded the Cretaceous diversity level (Thierstein and Berger, 1978; *Razvitie i Smena* ..., 1980, 1981; Shimanskii and Solov'ev, 1982; Benson *et al.*, 1984; Hansen, 1987; Keller, 1988). At the family level, the decline in the diversity of the main benthic groups was not as pronounced (see the table).

In general, the benthos composition in Paleogene seas was very similar to that during the Cretaceous. Only some groups, i.e., brachiopods, crinoids, and hydroids, became less common. Cephalopods lost their significance among the nektobenthos, while marine mammals became more numerous. The plankton composition was the same as in the Late Cretaceous.

Regarding the distribution of taxa and ethologotrophic groups (or guilds) in different parts of the sea, namely, in marginal coastal, relatively deep, and open shallow zones and areas of organic reefs, it is remarkable that in both the Cretaceous and Paleogene seas the taxonomic composition of communities differed in various biogeographic regions, most of all in the boreal and tropical, or equatorial, belts. It was rather impoverished in boreal seas, especially in shallow zones with organic reefs.

The marginal zones of Cretaceous basins with littoral and lagoon environments yielded communities dominated by bivalve mollusks and gastropods with less common ostracodes, foraminifers, and scaphopods. Brachiopods, serpulids, bryozoans, and nektobenthic cephalopods were of minor importance. The communities also contained numerous skeletonless worms and crustacean; thickets of aquatic vegetation, namely, green and red algae, occurred locally. Abundant epifaunal sestonophages were represented by free-lying, byssus, and anchored forms. Infaunal and semiinfaunal sestonophages as well as borers were relatively rare. Assorting epifaunal and infaunal detritophages and/or necrophages together with phytophages were numerous as well as nonassorting detritophages or silt feeders. Nektobenthic carnivores and/or necrophages were less common.

In addition to being dominated by bivalve mollusks, gastropods, and foraminifers, the lagoons of Paleogene boreal seas were inhabited by sponges, bryozoans, cirripeds, ostracodes, decapods, other crustaceans, and serpulids together with rare echinoids. Aquatic higher vegetation and algae were locally distributed. Dominant sestonophages were represented by epifaunal sessile (especially oyster) and anchored forms as well as infaunal and rare semiinfaunal organisms and borers. Numerous herbivores were associated with thickets of vegetation. Epifaunal and infaunal detritophages, necrophages, and carnivores were of minor importance. In the equatorial belt, the benthos was more diverse taxonomically in lagoons and bays with somewhat lower salinity. These areas were inhabited by bivalve mollusks, among which oysters were abundant, as well as gastropods, foraminifers (mainly miliolids), ostracodes, and other crustaceans (crabs, and others). Bryozoans, serpulids, and other worms were very characteristic here. Common thickets of algae and other marine vegetation occurred. Epifaunal and infaunal sestonophages dominated, though epi- and infaunal detritophages, including nonassorting forms, and herbivores were also

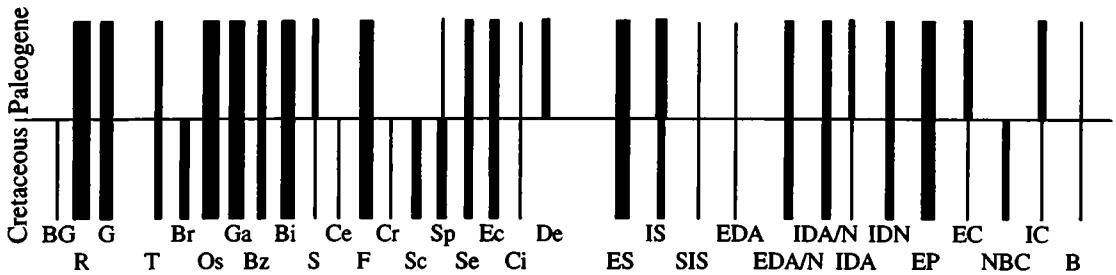

Fig. 1. Taxonomic and ethologo-trophic compositions of bottom communities from coastal zones of the Cretaceous and Paleogene seas.
Taxonomic groups: As – asteroids, BG – blue-green algae, Bi – bivalve mollusks, Br – brachiopods, Bz – bryozoans, Ce – cephalopods, Ch – chaetetids, Ci – cirripeds, Cr – crinoids, De – decapods, Ec – echinoids, F – foraminifers, G – green algae, Ga – gastropods, Hd – hydroids, Oph – ophiuroids, Os – ostracodes, R – red algae, S – scleractinians, Sc – scaphopods, Se – serpulids, Sp – sponges, St – stromatoporoids, T – traces.
Ethologo-trophic groups: B – borers, EC – epifaunal carnivores, EDA – epifaunal assorting detritophages, EDA/N – epifaunal assorting detritophages and necrophages, EP – epifaunal phytophages, ES – epifaunal sestonophages, IC – infaunal carnivores, IDA – infaunal assorting detritophages, IDA/N – infaunal assorting detritophages and necrophages, IDN – infaunal nonassorting detritophages, IS – infaunal sestonophages, NBC – nektobenthic carnivores, SIS – semiinfaunal sestonophages.

rather numerous. Semiinfaunal sestonophages and carnivores and/or necrophages were relatively scarce.

The coastal shallow zones of Cretaceous seas were predominantly inhabited by bivalve mollusks, gastropods, foraminifers, and ostracodes; brachiopods, echinoids, sponges, and bryozoans were less common. Scaphopods, serpulids, cirripeds, and crinoids occurred rarely; corals occurred even more rarely, and nektobenthic cephalopods were simply accessory. Abundant worms and crustaceans left numerous traces. Thickets of water vegetation, including green and red algae, were characteristic of these settings. Epifaunal sestonophages and phytophages dominated, whereas epi- and infaunal assorting detritophages and/or necrophages were the second most frequent groups. Infaunal sestonophages and nektobenthic carnivores were more abundant than semiinfaunal sestonophages and epi- and infaunal detritophages and carnivores (Fig. 1).

The coastal areas of Paleogene seas were inhabited by abundant bivalve mollusks, gastropods, and foraminifers (especially miliolids) along with common sponges, bryozoans, serpulids, ostracodes, decapods, and cirripeds. Echinoids were characteristic of equatorial seas and rare in boreal regions. Rich marine vegetation occurred locally. The ethologo-trophic composition of coastal communities was the same as that in lagoons (Fig. 1).

The Cretaceous communities of open shallow shelfs mainly yielded bivalve mollusks, gastropods, echinoids, brachiopods, sponges, foraminifers, and bryozoans. Scaphopods, decapods, and serpulids were of minor importance. Ostracodes were numerous only in boreal communities, whereas corals were widespread in tropical seas. Other groups, namely, crinoids, ophiuroids, asteroids, and cirripeds, were scarce. In both boreal and equatorial seas, deposits of this zone contained accumulations of red, green, and blue-green algae as well as numerous traces of worms and crustaceans. Nektobenthic cephalopods and fishes also occurred there. Against the background of dominant epifaunal sestonophages, including their free-lying, cemented, byssus, and swimming forms, the following representatives occurred: many assorting epi- and infaunal detritophages and/or necrophages; nonassorting infaunal detritophages and phytophages; and epi-, infaunal and nektobenthic carnivores. Infaunal sestonophages, including borers, were less common; semiinfaunal sestonophages and epifaunal assorting detritophages occurred as accessory forms (Fig. 2).

The shallow shelfs of Paleogene boreal seas were the habitats of bivalve mollusks, gastropods, and foraminifers associated with scaphopods, bryozoans, brachiopods, echinoids, crinoids, and various crustaceans and worms. Corals (scleractinians) were rare. Epifaunal, infaunal, and semiinfaunal sestonophages and borers composed the predominant ethologo-trophic group. Infaunal assorting and nonassorting detritophages as well as herbivores, carnivores, and/or necrophages were common forms, whereas epifaunal detritophages and parasites were scarce.

The communities of the shallow shelfs of Cretaceous tropical seas included diverse bivalve mollusks, gastropods, corals, foraminifers, and echinoids; common bryozoans, brachiopods, scaphopods, ostracodes, crabs and other crustaceans, serpulids, and skeletonless worms also occurred. Sponges, crinoids, and armored mollusks were rare. The aquatic vegetation comprised green and red algae and marine grass *Cymodocea*. Epi- and infaunal sestonophages, herbivores, and carnivores were characteristic of these communities. Infaunal detritophages, including nonassorting forms, were numerous. Epifaunal detritophages, semiinfaunal and boring sestonophages, and parasites also occurred there. Thus, both the taxonomic and the ethologo-trophic compositions of the Cretaceous communities in tropical shallow shelfs were very diverse (Fig. 2).

In shallow boreal and tropical zones with organic reefs, the taxonomic composition of Cretaceous commu-

nities was very different. Boreal seas were inhabited by bivalve mollusks (especially oysters) commonly associated with bryozoans, serpulids, cirripeds, sponges, foraminifers, gastropods, brachiopods, echinoids, and ostracodes; corals occurred only in the southernmost regions. Blue-green algae, which formed stromatolites, prevailed. Tropical seas were characterized by organic reefs composed of corals and rudists along with hydroids, stromatoporoids, chaetetids, and red and green algae. Many bivalve mollusks, bryozoans, foraminifers, calcareous sponges, brachiopods, serpulids, crinoids, gastropods, and echinoids inhabited the reefs. The ethologo-trophic composition of boreal and tropical communities was very similar: epifaunal sessile sestonophages and borers as well as phytophages were markedly predominant. Assorting epifaunal detritophages and/or necrophages, together with epifaunal and nektobenthic carnivores, were of secondary importance (Fig. 3).

In the shallow areas of Paleogene boreal seas, bryozoans and ahermatypic corals (scleractinians and octocorals) constructed the banks, which were also inhabited by serpulids, crinoids, cemented or anchored bivalve mollusks, brachiopods, hydroids, and calcareous sponges as well as gastropods, cirripeds and other crustaceous, echinoids, asteroids, and ophiuroids, which are all typical dwelles of organic reefs. In southern boreal regions, small constructions of coralline algae occurred, whereas in the north, the marine vegetation was represented by both higher plants and green algae. Sessile epifaunal sestonophages dominated in this zone. Anchored epi- and infaunal sestonophages, herbivores, and carnivores were less common here, and in- and epifaunal detritophages were scarce.

In Paleogene seas of the equatorial zone, where a vast expansion of reefs occurred during the Cretaceous, these morphostructures disappeared in the first half of the Paleocene. Small constructions were then formed mainly by coralline algae and, locally, by sponges and scleractinians (Bryan, 1991). True reefs composed of corals, coralline and green algae, bryozoans, certain foraminifers, bivalve mollusks, and gastropods appeared again only in the Oligocene. Epifaunal byssus and boring bivalve mollusks, herbivorous gastropods, echinoids, serpulids, sponges, solitary corals, ostracodes, crabs, and vagile foraminifers were associated with these reefs. Sessile sestonophages and microcarnivores (corals) prevailed. Other ethologo-trophic groups were common, namely, epifaunal mobile, anchored, and boring sestonophages; epifaunal detritophages and necrophages; carnivores; and herbivores;

Number of families of main benthic invertebrate groups from the Early Cretaceous to the Eocene

Benthic invertebrates	Cretaceous		Paleogene		
	Early	Late	Paleocene		Eocene
			Danian	Montian + Thanetian	
Foraminifers	24	47	45	46	55
Stromatoporoids	5		?	?	?
Sponges (as a whole)	90		45		
Sphinctozoans	5		?	?	?
Scleractinians	41	47	37	37	38
Hydroids (as a whole)	3	1	1	1	1
Sphaeractinida	1	1	–	–	–
Brachiopods	18	17	14		
Bryozoans	30	34	36	35 - 36	about 60
Gastropods	65	106 (93 in the Maestrichtian)	97	101	108
Bivalve mollusks (as a whole)	87	98 (86 in the Maestrichtian)	76	79	85
Hippuritida	6	6	?	1 genus	–
Scaphopods	2	2	2	2	2
Ostracodes	11	12	12	13	13
Echinoids (as a whole)	29	43 (38 in the Maestrichtian)	30	33	45
Hemicidaroida	2	2 (1 in the Maestrichtian)	–	–	–
Orthopsida	1	1	–	–	–
Clypeasteroida	–	?	?	1	6
Crinoids	20	17	?	9	
Asteroids	15	16	16	16	16

however, they had a secondary importance in the reef communities (Fig. 3).

In deep shelf zones, communities of both boreal and tropical Cretaceous seas included dominant bivalve mollusks, echinoids, foraminifers, sponges, and gastropods. Bryozoans, brachiopods, skeletonless organisms that left traces, and nektobenthic cephalopods were numerous here. Serpulids, crinoids, and ostracodes were less common and associated with scarce solitary corals, cirripeds, scaphopods, incrustated worms, asteroids, and ophiuroids. Free-lying, sessile, and swimming epifaunal sestonophages prevailed; infaunal and semiinfaunal sestonophages were of minor importance. Both assorting and nonassorting detritophages, especially infaunal forms, were the major components of these communities. Phytophages were rare, whereas nektobenthic, epifaunal, and infaunal carnivores occurred in all of the communities (Fig. 4).

The Paleogene communities of deep shelfs yielded siliceous sponges, small foraminifers, bryozoans, gastropods, bivalve mollusks (including many swimming forms), echinoids, and scaphopods. Solitary corals, brachiopods, ostracodes, serpulids, and crinoids were scarce. Epi- and infaunal sestonophages, infaunal detritophages, and epi- and infaunal carnivores dominated. Semiinfaunal sestonophages and epifaunal detritophages and/or necrophages were rare; phytophages were almost absent (Fig. 4).

In conclusion, we can state that the mass extinction and crisis of the biota at the Cretaceous–Paleogene boundary, which were major Phanerozoic phenomena, were neither short-term nor momentary events.

Most researchers now agree that the extinction at the end of the Cretaceous was long, stepwise, and asynchronous for different groups. Most likely, there were several factors that caused the extinction. Including the possibility of an impact event, among the factors that caused the general destabilization of the environments there could be the following: an intensification of volcanism and associated phenomena; a reduction of epicontinental sea areas; a general cooling followed by a short warming; an oxygen deficiency; and, finally, a decrease in the nutritient input from land areas. Certain groups were influenced by different factors: planktonic foraminifers and calcareous nannoplankton were affected by the lower intensity of sun radiation, changes in the chemical composition and temperature of water, and the decline in nutrition. Benthic forms, namely, brachiopods, bivalve mollusks, and bryozoans, were affected by the nutritient deficiency, the charge from carbonate oozes to an unstable clayey substratum, the contraction of habitats, the oxygen deficiency, the cooling, and other factors.

Therefore, all components of the biota were influenced by a combination of abiotic and biotic factors, and changes in one group of organisms affected, as

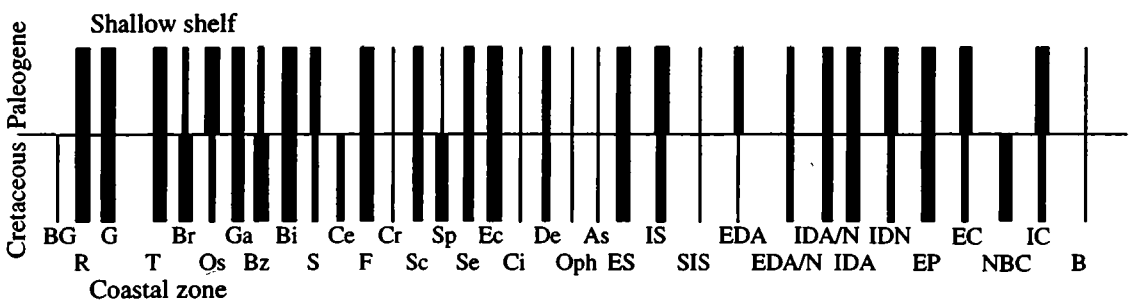
Shallow shelf
Paleogene
Cretaceous
BG G Br Ga Bi Ce Cr Sp Ec De As IS EDA IDA/N IDN EC IC
R T Os Bz S F Sc Se Ci Oph ES SIS EDA/N IDA EP NBC B
Coastal zone

Fig. 2. Taxonomic and ethologo-trophic compositions of bottom communities from the shallow shelf zones of the Cretaceous and Paleogene seas.
See Fig. 1 for explanations.

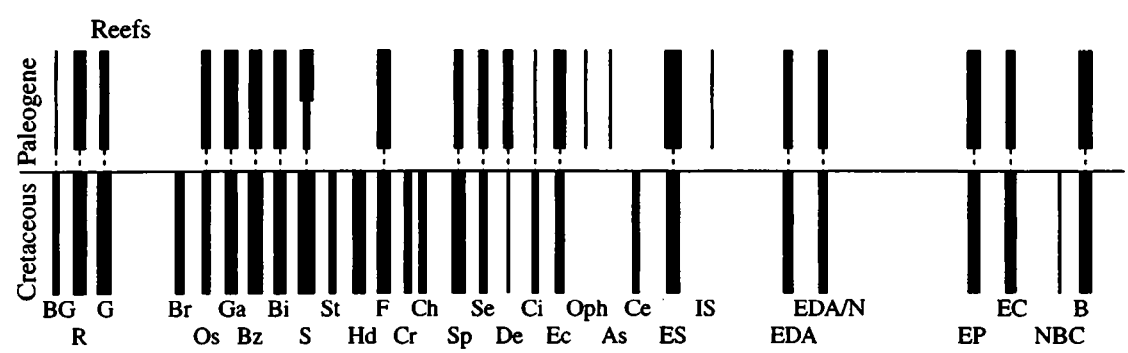

Fig. 3. Taxonomic and ethologo-trophic compositions of reef communities from shallow zones of the Cretaceous and Paleogene tropical seas.
See Fig. 1 for explanations.

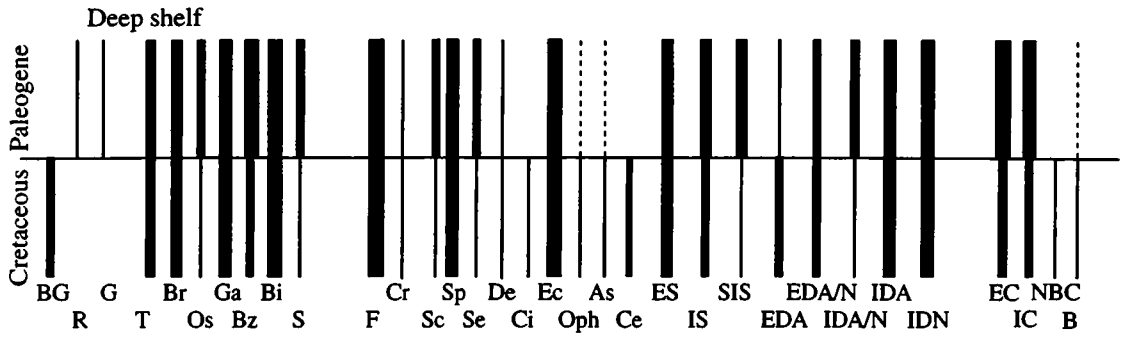

Fig. 4. Taxonomic and ethologo-trophic compositions of bottom communities in the deep shelf zones of the Cretaceous and Paleogene seas.
See Fig. 1 for explanations.

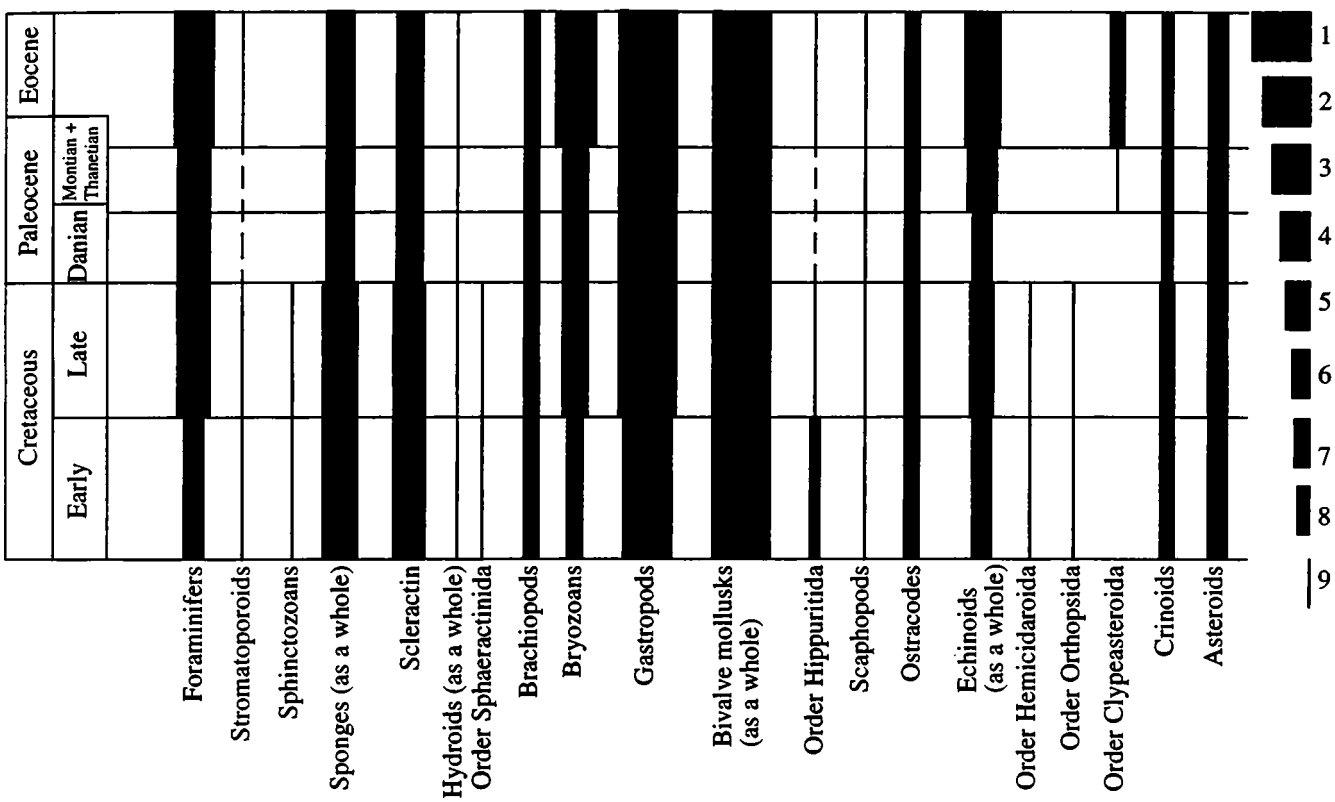

Fig. 5. Distribution of main benthic invertebrate groups composing the Cretaceous and Paleogene shelf communities.
1 - 9 – number of families: 1 – more than 81; 2 – 61 - 80; 3 – 51 - 60; 4 – 41 - 50; 5 – 31 - 40; 6 – 21 - 30; 7 – 11 - 20; 8 – 6 - 10; 9 – 1 - 5.

a consequence of coevolutionary processes, other groups and the biota as a whole.

All the changes were initiated by an abrupt productivity decline in ocean waters due to the rapid extinction of pelagic plankton, which may have resulted from both extraterrestrial and terrestrial events.

However, the catastrophic abiotic events were superimposed on the rearrangement of biota (including the marine groups), which had already occurred at the end of the Cretaceous and was caused by the reduced diversity of some planktonic, nektonic, and benthic groups. Groups were rearranged differently depending on the internal evolutionary factors specific to each group.

The rearrangement of biota, which began no later than in the early Maastrichtian, was only reinforced and hastened, according to Hoffman (1986), by the "terminal Cretaceous event."

Despite the substantial renewal of benthos at the Cretaceous–Paleogene boundary due to the intense extinction and the subsequent appearance of new taxa, the main transformations occurred at the species, generic, and, to a lesser extent, family levels (see the table), but they barely affected higher taxa, as was the case, for example, at the Permian–Triassic boundary. Only four orders, three of which included only one family each at the end of the Cretaceous, became extinct,

namely, Sphaeractinida among hydroids and Hemicidaroida and Orthopsida among regular echinoids. Only the order Hippuritida (bivalve mollusks), composed of six families, was of considerable importance in the biota of Cretaceous equatorial seas, resulting in the fact that its extinction greatly influenced the composition of the reef community. The renewal of benthos due to the appearance of high taxa was even less substantial: only one order of irregular echinoids, Clypeasteroida, appeared; the data concerning the first occurrence of this order is controversial and varies from the end of the Cretaceous to the middle Paleocene (Fig. 5).

The ethologo-trophic composition of bottom communities remained almost the same. Only taxa of the generic and family levels composing one or another ecologic group experienced changes. This was also true for reefs, which, as stated above, suffered greatly during the crisis and almost disappeared at the beginning of the Paleocene. Their complete reestablishment occurred only in the Oligocene. The main reef builders, excluding rudists, remained the same, although hydroids, chaetetids, and stromatoporoids lost their importance as frame builders among the subordinate members of communities. At the same time, crinoids and brachiopods, the Cretaceous reefophilous organisms migrated to deeper zones.

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Dynamic Relationships between Principal Tectonic Events Related to the Earliest Evolution of the North Atlantic, Mediterranean, and Urals–Okhotsk Mobile Belts

S. G. Rudakov

Geology Department, Moscow State University, Moscow, 119899 Russia

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Abstract – Principal tectonic events of the Baikalian stage recognized in the Mediterranean mobile belt also affected the North Atlantic and Urals–Okhotsk belts. Major structural changes related to the Dalslandian (Grenvillian), late Dalslandian (Dehli, after Khain), Cadomian, and Salair pulses of tectogenesis along with those that occurred in the Vendian are compared for all the belts. The first two pulses reflect progress in the evolution of the belts, which is evident from the opening of ocean basins under conditions of general crustal extension and decomposition of the pre-Late Proterozoic Pangea-I supercontinent. The Dalslandian (1200 - 1000 Ma) events were monotonous, but after the late Dalslandian tectogenesis (850 - 800 Ma), the belts developed differently. The transition from rifting to spreading stages was faster in the Mediterranean belt, where the Prototethys Ocean opened earlier than others. The Vendian was a period of the maximum opening of ocean basins. Their gradual closing started with the Cadomian tectogenesis (about 650 - 570 Ma), and local collisions of this time are recorded in all the belts. The Prototethys, with its vast peri-Gondwanian part, experienced maximum reduction. The Salair tectogenesis (520 - 490 Ma) was clearly manifested by diverse transformations in all belts. Many basins became closed at that time, but in some areas the collision was local, and a secondary opening of oceans occurred in the peri-Gondwanian part of the Mediterranean belt and, probably, in the Urals. Alternative tectonic processes of complementary geodynamic significance were typical of the Cadomian and Salair events. It is concluded that in the Late Proterozoic–Early Paleozoic the North Atlantic, Mediterranean, and Urals–Okhotsk mobile belts represented an integral global system, whose separate parts evolved under the influence of a universal tectonic rhythm.

Key words: *mobile belts (Prototethys, Urals–Okhotsk, North Atlantic), Late Proterozoic, beginning of the Paleogene, correlation.*

INTRODUCTION

Intensive tectonic events during the earliest evolutionary stages of the North Atlantic, Mediterranean, and Urals–Okhotsk mobile belts, which formed a global system of linear structures and split the Late Proterozoic supercontinent Pangea-I into major continental plates, were very significant in the history of crustal evolution. I showed earlier that the most important events of the earliest tectonic evolution of the Mediterranean occurred about 1100 - 1000 and 850 - 800 Ma ago as well as during the Cadomian and Salair tectogeneses (Rudakov, 1991). The two former time intervals correspond to periods of ocean-basin opening, whereas the Cadomian and Salair tectogeneses are related to the terminal Baikalian phase of mobile belt evolution. Similar events in the North Atlantic belt are correlative with those in the two belts bordering the eastern part of the European continent (Rudakov, 1993).

One of the reasons for including the Urals–Okhotsk mobile belt in this analysis is the fact that the Urals fragment belongs to the eastern borderland of Eastern Europe. Another reason is related to global reconstructions (Khain and Bozhko, 1988), which suggest that direct relationships existed between the Prototethys and the Kazakhstan–Tien Shan basin of the Urals–Okhotsk belt. These same reconstructions revealed that the latter basin is connected with the Indo-China fragment of the Mediterranean belt, which also makes it reasonable to analyze the earliest events of the Indo-China evolution. Thus, the main purposes of this investigation are, first, to correlate the principal tectonic changes in the evolution of the mobile belts during the aforementioned and Vendian intervals, and, second, to reveal the geodynamic relations between these changes.

These problems are crucial because our knowledge on the earliest evolution of the mobile belts is insufficient in comparison with the subsequent stages in their development. The results obtained (Fig. 1), although schematic, characterize the principal features of the North Atlantic, Mediterranean, and Urals–Okhotsk belts in the Late Proterozoic–Earliest Paleozoic.

DALSLANDIAN TECTONIC EVENTS

Isotopic ages from 1200 (often from 1100) to 1000 Ma mark tectonic events that occurred between the Middle and Late Riphean in the evolution of all the mobile belts analyzed.

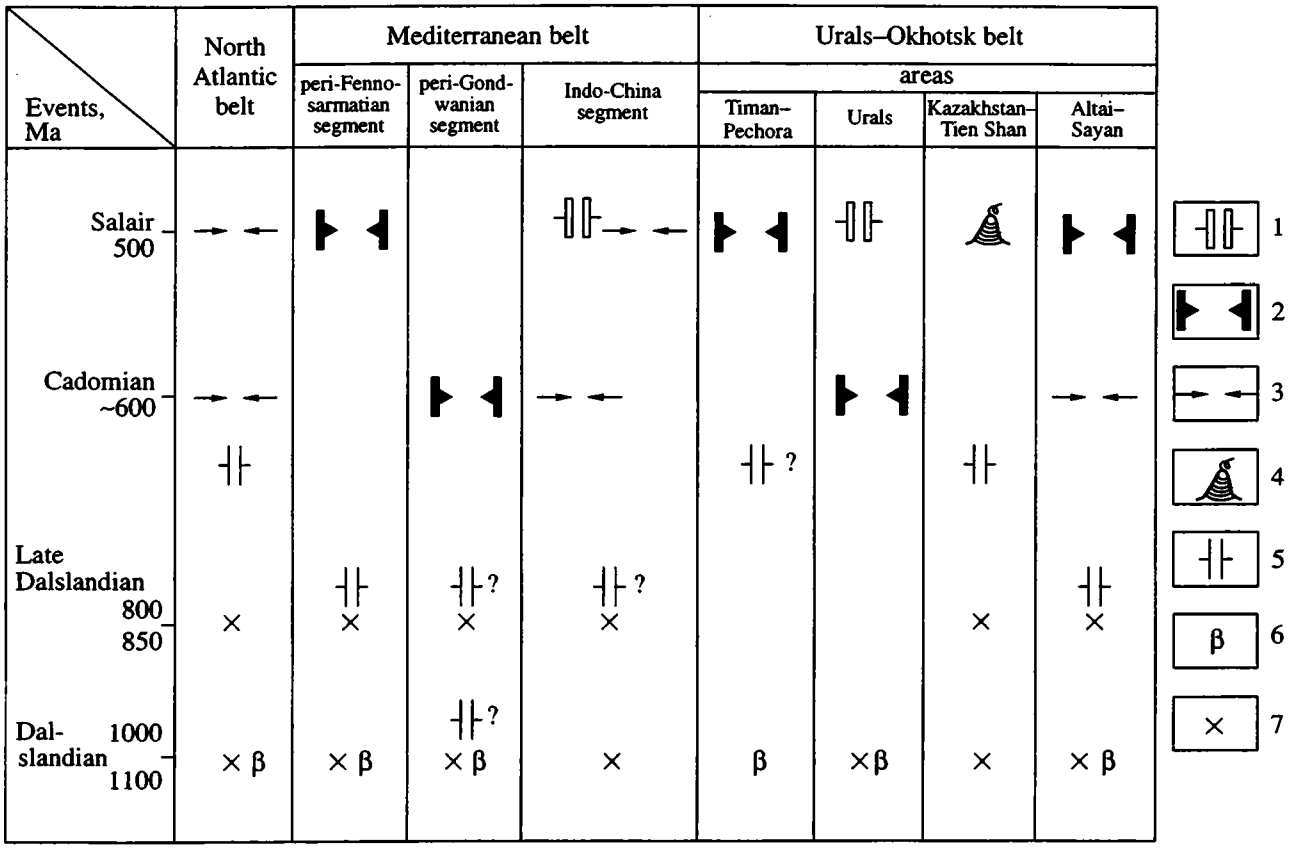

Fig. 1. Correlation of major tectonic events in the evolution of the North Atlantic, Mediterranean, and Urals–Okhotsk mobile belts in the Late Proterozoic–Earliest Paleozoic.

1 – regeneration of spreading; 2 – complete basin closings; 3 – local collision with local basin closings; 4 – island-arc magmatism; 5 – initiation of spreading (as indicated by ages of oldest oceanic deposits); 6 – youngest manifestations of initial basite magmatism; 7 – Dalslandian and late Dalslandian granitoids.

Traces of the Dalslandian (Grenvillian) tectogenesis, which resulted in the formation of the North Atlantic belt that separated North America and Eurasia, are distinguished in the both of these continents. In North America, these traces comprise the entire complex of the Grenvillian events (s.s.); peaks of metamorphism and deformation occurred about 1100 and 950 Ma ago. The granitoids in Scotland (1020 Ma old) and Ireland (1070 Ma old) are related to the same impulse of tectogenesis. In the European borderland of an incipient mobile belt, this tectogenesis resulted in the deep reworking of the ancient rock complexes in the Sveconorwegian area and the Pentevrian complexes in the Armorican massif (Khain and Bozhko, 1988). Events that reflect an intensification of continental rifting have similar datings. These events were responsible, in particular, for the formation of basite dike swarms in Greenland (988 Ma), Sweden (980 - 950 Ma), and Finland (about 1050 Ma) (Pesonen *et al.*, 1989).

Active continental rifting, which resulted in the opening of the ocean basin that separated Laurasia from Gondwana, is also notable in the Mediterranean mobile belt. This tectonic activity caused the formation of basite complexes in the adjacent platform margins and within the belt. Gabbro-dolerites of the Volyno-Podolia region (1060 ± 41 and 1041 ± 28 Ma) can serve as an example for the pericratonic complex. Within the belt, orthoamphibolites of the Pennine Alps (1020 Ma) and the northern Dinarides (1010 Ma) are of the same tectonic origin.

In addition, granitoids and gneisses of some pre-Baikalian structures, whose relics are preserved in the mobile belt, have similar ages, which can be exampli-fied by rocks from the Pannonian basin basement (1180 - 975 Ma; Rudakov, 1985) and by the Prilep granitoids in the Pelagonian massif (about 1000 Ma; Papanikolaou and Stojanov, 1983). In the history of Indochina, tectonic events that occurred 1000 Ma ago marked the beginning of the final consolidation of the South China platform basement and, in addition, were accompanied by the gradual opening of the Viet–Laos basin (a part of the future Prototethys), which had been initiated as long ago as in the middle Riphean (Khain, 1984).

The Dalslandian events are also clearly manifested in the evolution of the Urals–Okhotsk heterogeneous mobile belt. In the Timan–Pechora–Barents Sea region, the age of the Barmin-Chernoretski and South-Timan gabbro-diabase complexes is about 1000 Ma (Belyakova and Stepanenko, 1991). In the Urals fragment of the belt, about 1100 - 1000 Ma ago, the Dalslandian tectogenesis resulted in the formation of paired metamorphic belts, folding, and granitization (Lennykh, 1984). Datings of 1069 - 987 Ma, obtained for gneisses and granitoids from the basement of the West Siberian plate, indicate the same diastrophism (Surkov and Zhero, 1981). In addition, similar ages were determined for amphibolites of the Polar Urals, whose chemical composition is similar to that of continental tholeiites. Lennykh (1984) noted that the detachment of microcontinents and the formation of marginal seas directly followed the folding and metamorphism on the western slope of the Urals. Khain *et al.* (1993) showed that the Grenvillian granitoids of Taimyr, the Enisei Ridge, and the northwestern part of the East Sayan suggest that the boundary between the Siberian paleocontinent and the forming ocean appeared before the beginning of the Late Riphean. Granitoids of the Teskhem–Mugur complex of the Sangilen microcontinent, for which a U-Pb isochrone age of 1080 Ma was obtained for zircons, seem to represent the same group of intrusions in the Altai–Sayan region (Gibsher *et al.*, 1991). The Riphean ophiolitic belt, distinguished by Khain *et al.* (1993), marks one of the oldest basins in the expanding Central Asian ocean.

In the history of the Kazakhstan–Tien Shan area of the Urals–Okhotsk belt, the Icedonian folding that occurred 1100 Ma ago was accompanied by granitization and regional metamorphism (Zaitsev, 1984). Synorogenic granitoids of Kazakhstan were dated as 1200 and 1100 Ma, whereas those of northern and central Tien Shan have ages of 1270 and 1070 Ma (Khain and Bozhko, 1988). The same tectogenesis resulted in the closing of some aulacogens in the northern periphery of the North China platform, which was accompanied in places by folding and low-grade metamorphism (Milanovski, 1991).

In the history of the Altai–Sayan area and adjacent platform margins, the Dalslandian tectonic events caused a structural reorganization of the pre-Baikalian basement in the southern part of the Siberian craton and formed the granitoids of the Barguzin complex (1014 Ma) and the migmatites of northern Mongolia (1058 - 950 Ma) (Khain and Bozhko, 1988). Postel'nikov and Museibov (1992) noted that substantial rifting occurred in the southwestern part of the Siberian platform during the Grenvillian phase, 1100 - 1000 Ma ago. In the Altai–Sayan region proper, the same tectogenesis resulted in the formation of complex diabase intrusions dated as old as 1190 Ma (Altukhov, 1986).

Thus, a phase of tectogenesis with folding, granitization, and metamorphism, manifested in all the belts and along the periphery of adjacent cratons, is a common feature of the Dalslandian or Grenvillian events. In addition, the same interval between 1200 (1100) and 1000 Ma comprised events resulting in the destruction of the oldest continental crust. Basite magmatism, registered in all the belts, was the most characteristic of such events, although it was separated from the initial oceanic sedimentation by different time spans. These competitive processes of crustal evolution, which are closely related in time, seem indicative of the rifting stage that preceded the opening of ocean basins in the mobile belts. During the interval 1200 - 1000 Ma, compressive deformations and related processes were abruptly and repeatedly replaced by extension. The obvious synchronism of events in all the belts may indicate, to a certain extent, that the integrity of the Pangea-I supercontinent had not yet been lost. This is also confirmed by paleomagnetic data (Pesonen *et al.*, 1989), according to which the Grenvillian and Sveconorwegian provinces were apparently parts of a single continuous belt during the Dalslandian tectogenesis. The same data indicate that a considerable transposition of the belt fragments began about 1000 - 900 Ma ago. Rifting in the belts continued and became even more intensive after the Dalslandian events.

LATE DALSLANDIAN (DELHI) TECTONIC EVENTS

Tectonic events that occurred at about 850 - 800 Ma are also distinguishable in the history of the analyzed mobile belts. In the North Atlantic belt, where rifting was in progress, they mark the Bagko complex in Norway and the Torridon complex in Scotland, which are as old as 825 Ma and 810 ± 17 Ma, respectively (Stewart, 1982). Within the Mediterranean belt, the Prototethys Ocean opened soon after the period 850 - 800 Ma and separated Laurasia from the African part of Gondwana (Fig. 2A). Shortly thereafter, volcanism of ensialic and ensimatic island arcs supplemented the terrestrial volcanism here (Rudakov, 1991). Undoubtedly, the Arabian paleoocean was also one of the elements of the Prototethys. Recent publications (Abdelsalam and Stern, 1993; Kroner *et al.*, 1992) show that within this structure oceanic magmatism was initiated after 900 Ma, whereas its island-arc complexes originated in the late Riphean and Vendian.

Approximately during the time under discussion, in the Indo-China fragment of the Mediterranean basin, the Viet–Laos basin with oceanic crust opened between the South China and Indo-China continental blocks, as is evident from ophiolite occurrences (837 Ma) in the Guansi Province in the extreme southern part of China (Zhai *et al.*, 1985). The Yangtze orogenic events occurred at about 850 Ma as well and were terminal in the basement consolidation of the South China craton, which bordered the newly formed ocean basin of the Prototethys.

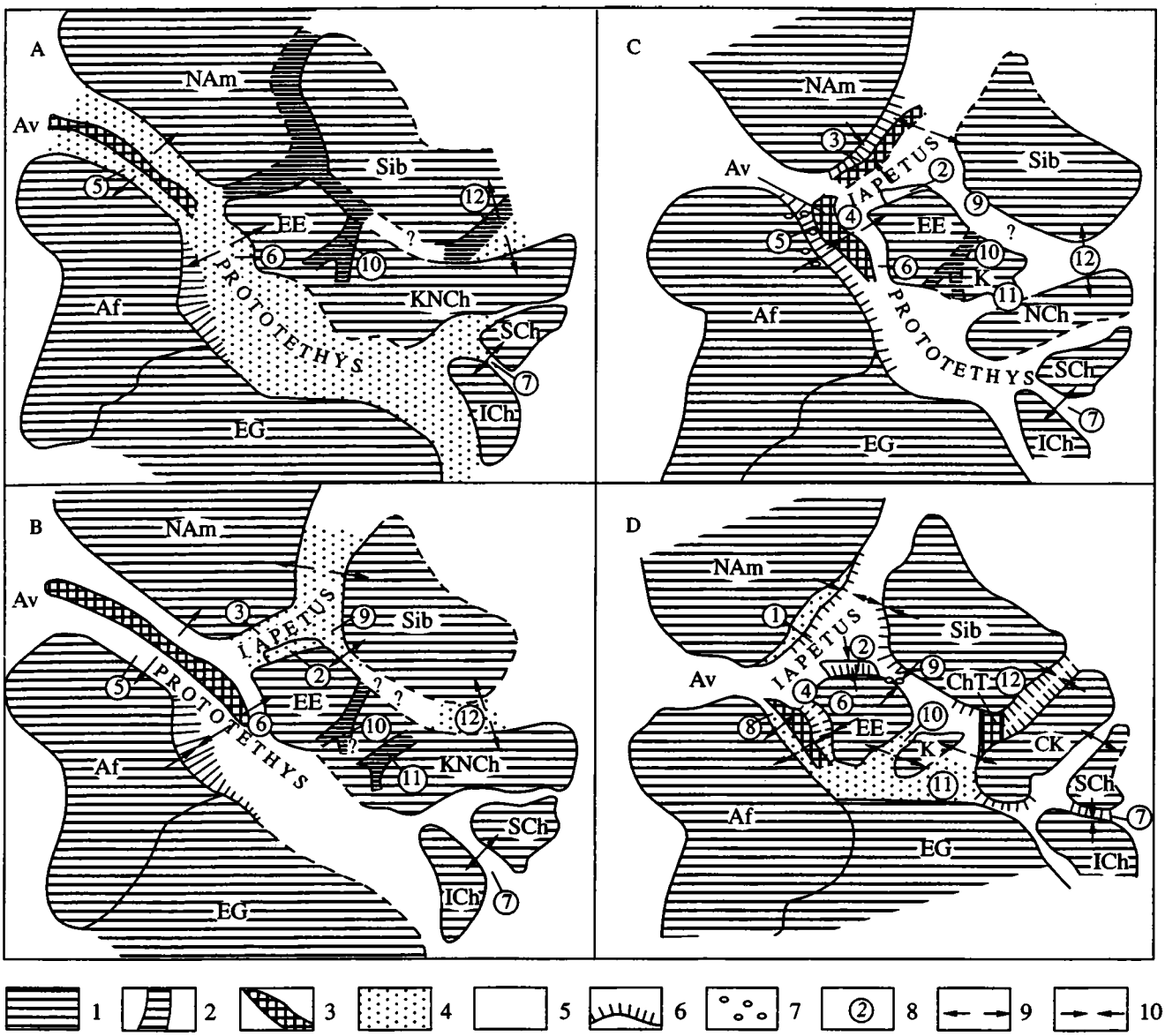

Fig. 2. Schemes of principal changes in the structure of the North Atlantic, Mediterranean, and Urals–Okhotsk mobile belts. A – after the late Dalslandian events; B – during the Vendian (prior to the Cadomian tectogenesis); C – after the Cadomian (the Vendian–Cambrian boundary period); D – after the Salair events (the Cambrian–Ordovocian boundary period). Schemes are based on reconstructions for the Precambrian after Khain and Bozhko (1988). 1 – major continental blocks; 2 – zones of continental rifting; 3 – major island arcs; 4 – newly formed ocean basins; 5 – young ocean basins; 6 – closed ocean basins (accretional complexes); 7 – areas of orogeny; 8 – numbers of structural units; 9 - 10 – conjugated processes: 9 – tension, 10 – compression. Numbers in circles: 1 - 4 – structural units of the North Atlantic belt: 1 – Grampianides, 2 – Caledonides of Scandinavia, 3 – the marginal Monian Sea, 4 – the Tornquist Sea; 5 - 8 – structural units of the Mediterranean belt: 5 – peri-Gondwanian (Cadomian) basin, 6 – peri-Fennosarmatian basin, 7 – Viet–Laos basin, 8 – the earliest basins of the Paleotethys; 9 - 12 – structural units of the Urals–Okhotsk belt: 9 – Timan–Pechora–Barents Sea area, 10 – Uralian region, 11 – Kazakhstan–Tien Shan area, 12 – Altai–Sayan area.

Continental blocks: Af – African, EG – East Gondwanian, EE – East European, ICh – Indo-China, K – Kazakhstan, KNCh – Kazakhstan-North China, NAm – North American, NCh – North China, Sib – Siberian, SCh – South China (Yangtze); island arcs: Av – Avalonian, ChT – Chengiz–Tarbagatai.

Within the Urals–Okhotsk belt in Kazakhstan, volcanism was active 850 - 800 Ma ago and associated with tensional deformation and gradual destruction of the continental crust. During the same late Dalslandian period, granitoids intruded the western periphery of the Siberian craton, for example, the Taimyr (850 ± 50 Ma) and the Enisei Ridge (850 - 820 Ma) (Khain *et al.*, 1993). At about 850 - 800 Ma, tectogenesis affected the southern part of the Siberian and northern China platform structures that bordered the Altai–Sayan area, where, according to recently obtained data, an ocean basin had existed since the Riphean. Khain (1991) believed that this

primitive ocean was formed as a result of the destruction of the older continental crust of the eastern Sayan and northern Mongolia. Volcanic island arcs formed here at the end of the Late Riphean.

In summary, I should note that one of the significant terminal impulses of tectonic rifting during the long evolution of the mobile belts seems to be related to the late Dalslandian (Delhi) epoch (850 - 800 Ma). Judging from the formational character of rock complexes, this epoch marked the terminal rifting and initial spreading in the European fragment of the Mediterranean belt. The openings of the Iapetus ocean in the North Atlantic belt and a similar basin in the Timan–Pechora–Barents Sea area were imminent at this time. Paleomagnetic data indicating the commencement of Baltic shield motion northward after the interval 850 - 800 Ma also confirm this inference (Shipunov and Chumakov, 1991). In the Altai–Sayan area of the Urals–Okhotsk belt, the transition to spreading started in the Riphean. The lack of late Dalslandian events in the Urals may indicate that the tectonic regime was stable there. The increasing destruction of continental crust, which resulted in the Pangea-I breakup into northern (Laurasian) and southern groups of continents, was typical of all mobile belts. The first group incorporated the North American, East European, and Siberian plates, whereas the second group comprised the continental blocks of Gondwana.

THE VENDIAN PERIOD

The Vendian tectonic events in the analyzed mobile belts are also closely interrelated in a dynamic aspect (Fig. 2B). The separation of the Eurasian plate from the North American plate and the opening of the Iapetus ocean in the North Atlantic belt seem to have been related to the formation of the Timan–Pechora–Barents Sea structural area on the northern edge of Eastern Europe, which linked the Iapetus with the gradually expanding Urals–Okhotsk belt. The opening of the Iapetus is confirmed by paleomagnetic data (Shipunov and Chumakov, 1991) indicating that the Baltic shield drifted southward along with Eastern Europe. In the Iapetus, the earliest island-arc complexes of Wales, Malvern, and other areas are 700 - 640 Ma old (Thorpe *et al.*, 1984). In the Timan–Pechora region, Belyakova and Stepanenko (1991) also noted different relics of a mature volcanic arc, which probably extended into the Telpos zone of the pre-Uralides in the Polar Urals. These authors speculate that basins with oceanic crust might exist northeast of the Timan–Pechora region.

The separation of the Siberian and North China plates resulted in the opening of the Central Asian ocean in the Urals–Okhotsk belt. This event is best constrained by rock ages reported from the Altai–Sayan area, where the mass appearance of ophiolites indicate the beginning of the spreading period, which was initiated in the late Riphean.

CADOMIAN TECTONIC EVENTS

Tectonic events at the Vendian–Cambrian boundary appeared to be the most important in the evolution of the Iapetus and Prototethys; the Avalon island arc, whose parts collided with North America and North Africa, played a decisive role here (Fig. 2C). The first collision accomplished the evolution of the marginal Monian Sea and resulted in the metamorphism of the corresponding rock complex, which was unconformably overlain by Cambrian rocks. In western Gondwana, the Cadomian tectogenesis was worthy of the title "Pan-African," as it was registered throughout the northern part of this continent. The newly formed continental crust was accreted to the periphery of the Arabian-Nubian shield, as was recently reported (Sultan *et al.*, 1992). The Cadomian tectogenesis is also clearly recognized in many regions of central and western Europe, which presumably were of Gondwanian origin (Khain and Rudakov, 1991; Erdtmann, 1991). The closing of initial basins in the peri-Gondwanian part of the Prototethys, its attachment to the North African margin of Gondwana, and the subsequent rapid transformation of the region into a young orogen were the principal tectogenic results in the Mediterranean belt. In the basins of the opposite, peri-Fennosarmatian part of the belt, which were not subjected to Cadomian tectogenesis, the previous tectonic regime remained stable.

The Cadomian events responsible for the closing of the initial basins, folding, and metamorphism also occurred in some Indo-China segments of the belt, where the China–Burma massif was coupled with Gondwana (Khain, 1984). These events clearly resemble changes in the evolution of peri-Gondwanian Europe in the Prototethyan regions (Rudakov, 1991); although, in eastern Gondwana, the Cadomian tectogenesis left only local traces, in contrast to its western parts. Puchkov (1993) was correct in relating the Cadomian tectogenesis with the closure of the Riphean ocean in the Urals, the opening period of which is not yet clear. The epi-Cadomian orogeny is well-pronounced here, as is indicated by molassoids of the Ashinian Group. The closing of the West Asian Prototethys and subsequent Cadomian orogeny in the region were apparently related to intensive spreading in the adjacent Central Asian ocean, which also caused the opening of new basins with oceanic crust. In particular, they are the Vendian–early Cambrian ophiolites in the Tien Shan and Kazakhstan, (Kheraskova and Dergunov, 1986), though they differ from the true oceanic type (Khain and Seslavinskii, 1991) and originated in pull-apart structures with limited spreading (Zaitsev, 1984). The specific character of the Vendian ophiolites of Kazakhstan seems to indicate that their root zones were affiliated with the periphery of the gradually expanding Central Asian ocean, which probably had connections with the relic epi-Cadomian basin of the Prototethys through the Kazakhstan–Tien Shan region.

The Cadomian tectogenesis was also manifested locally in the Altai–Sayan segment of the Central Asian ocean. Here, within the eastern Sayan and northern Mongolia, the tectonic imbrication of island-arcs and some oceanic structures and their attachment to the Siberian continent commenced 640 - 620 Ma ago along with metamorphism of glaucophane-schist facies (Postnikov *et al.*, 1991). Khain (1991) emphasized the intraoceanic nature of these events and the further persistence of the ocean basin.

Thus, during the Cadomian epoch, the closings of some basins had dynamic relations with the openings of others. This situation remained stable, and the progressive extension in basins that bordered continents of the "northern" group was caused by events of spreading inside these continental masses.

SALAIR EVENTS

Many of the principal changes in the evolution of all the analyzed mobile belts were related to the Salair tectogenesis that occurred in the second half of the Cambrian and during the initial Ordovician (Fig. 2D). The Iapetus history of that time corresponds to the closing of peripheral basins adjacent to the North American and East European continents. This is exemplified by the Grampian basin deposits preserved in Scotland. On the East European side, simultaneous compressional deformations resulted in the Finnmark tectogenesis, which affected the Scandinavian Caledonides, as well as in volcano-plutonic activity and ophiolite obduction (Khain and Seslavinskii, 1991; Anderton, 1982). The age of the corresponding metamorphism is 530 ± 35 Ma. Požaryski (1990) also recognized Grampian events in some other segments of the Iapetus. These are represented by the South Jutlandian, Lysogursky, and Malopolsky tectonic terranes, which were initially located in close proximity to Eastern Europe. The Salair events in the Iapetus also affected the Avalonian arc, in particular, the Brabant massif, which, judging by acritarchs, is barren of uppermost Middle Cambrian–lowermost Tremadocian deposits (Vanguestaine, 1991).

Gradually increasing regressive changes in the evolution of the "Tornquist Sea," which was a large strait connecting the Iapetus and Prototethys, were also related to the Late Cambrian tectogenesis (Rudakov, 1993). Paris and Robardet (1990) and Torsvik *et al.* (1991) emphasized the considerable decrease in fauna variations at the beginning of the Ordovician, and this was used in the identification of the Tornquist Sea (Cocks and Fortey, 1982). The sea disappeared during the Latest Ordovician–Early Silurian, continuing a gradual closure of the Iapetus that lasted through a series of tectogenic phases from the Cadomian to the Calescandian (Earliest Devonian) (Khain and Seslavinskii, 1991).

In the Uralian segment of the Urals–Okhotsk belt, which was an eastern borderland of Eastern Europe, an ocean basin that was probably not inherited from the Riphean structure existed (Puchkov, 1993). This paleocean connected the Urals–Okhotsk and Mediterranean mobile belts. Dynamic relationships between the Iapetus reduction and the opening of the Rheicum ocean in Central Europe, which was caused by crustal extension, is quite obvious (Cocks and Fortey, 1982) and is confirmed by the development here of mafic-ultramafic and bimodal complexes of Late Cambrian–Early Ordovician age (Erdtmann, 1991; Pin, 1990). The gradually progressing destruction of the crust resulted in the formation of many continental microplates, which collided within the basin that opened during the early Paleozoic.

Adjacent areas of the Mediterranean belt were affected by conjugated, though different processes. Compressional deformations closed the peri-Fennosarmatian relic basin of the Prototethys bordering Eastern Europe from the southwest (Rudakov, 1991). The eastward extension of this basin can be assumed from the presence of similar Uppermost Proterozoic–Lowermost Paleozoic complexes in some areas of the southern borderland of the East European craton (Rudakov, 1991). Nearly simultaneously, a basin in the peri-Gondwanian part of the belt, which later formed the Paleotethys, was in its initial stage of opening. In the Indo-China segment of the Prototethys, the Viet–Laos basin evolved continuously from the second half of the Late Riphean through the Early Cambrian. The basin is distinguishable due to ophiolites that are probably Vendian–Early Cambrian in age. It closed during the Salair events (Khain, 1984). The Late Cambrian tensional deformation (the Yunnan phase) also affected the South China platform margin (Khain and Seslavinskii, 1991).

Salair tectonic processes of a different significance were characteristic of the extensive Urals–Okhotsk mobile belt. The Timan–Pechora–Barents Sea area closed here in pre-Ordovician time (Belyakova and Stepanenko, 1991) and was later involved in orogeny.[1] The turning point in the evolution of this area, which connected the Urals–Okhotsk and North Atlantic oceanic areas, was related to the closing of the latter and the formation of the Scandinavian Caledonides, which are the oldest in Europe.

For the Urals segment of the Urals–Okhotsk belt, I should emphasize that (1) the epi-Cadomian orogenic phase appeared to be very short, and (2) in the second half of the Cambrian, the Salair tectogenesis (Klyuzhina, 1984) resulted in the reopening of an ocean basin here, which connected the Urals–Okhotsk and Mediterranean mobile belts.

In the history of the Altai–Sayan region, the Salair tectogenesis was responsible for the principal events

[1] Getsen (1991) reported that the structure of Upper Proterozoic deposits in this region was formed at the end of the Vendian. Nevertheless, the similarity of some Upper Proterozoic–Cambrian magmatic formations (Belyakova and Stepanenko, 1991) and the nearly complete absence of Cambrian deposits seam to correspond more to the Salair epoch for the closing of the basin.

during the closing of the ocean basin. The sequence of events included the obduction of ophiolites and related complexes onto the Siberian continental margin, the attachment of microcontinents to it, and the subsequent formation of granite-gneiss domes (Khain, 1991). The Salair tectogenesis affected the evolution of all the mobile belts and provides a good opportunity for revealing the conjugation of events. Changes caused by the considerable migration of the East European plate relative to adjacent continental blocks, which, according to paleomagnetic data (Torsvik *et al.*, 1991), also rotated, are easily recognized. The convergence of Eastern Europe and northern Gondwana resulted in the closing of the peri-Fennosarmatian relic basin of the Prototethys, which was compensated, to a certain extent, by the opening of the Paleotethys. The same plate motions enhanced the Iapetus reduction and the closing of the Timan–Pechora–Barents Sea basin, located on the northern edge of Eastern Europe. Thus, the Salair events in the northeastern, northwestern, and southwestern borderlands of the East European craton reflected a regime of intense compression that existed there and resulted in the Iapetus reduction and the complete closing of the Prototethys.

The opening of the Uralian ocean within the Urals–Okhotsk belt connected it with the Kazakhstan–Tien Shan basin of the Central Asian ocean and supported the existence of the latter. The Salair events in this basin resulted in the formation of the Chingiz–Tarbagatai island arc bordering the Dzhungar–Balkhash area of the basin in the east, where the preceding regime remained unchanged during the Ordovician. In the Altai–Sayan area of the Central Asian ocean, located east of the newly formed arc, drastic transformations occurred during the second half of the Cambrian and at the beginning of the Ordovician and resulted in the closing the ocean basin[2] due to sialic crust imbrication and the coupling of newly formed crustal zones with those consolidated earlier. For instance, the Central Mongolia massif was integrated with the southern part of the Siberian continent. Later, after a short orogenic phase, rifting resumed in the Eastern Sayan (Khain, 1991).

The Salair events in eastern Asia were characterized by a prevalence of tectonic compression; although, in places such as the Zinlin ocean between the North and South China continental blocks, ocean basins remained open (Khain and Seslavinskii, 1991).

CONCLUSION

The conducted analysis allows us to characterize the general features of evolution of the North Atlantic, Mediterranean, and Urals–Okhotsk mobile belts during the Late Proterozoic and Earliest Paleozoic epoches.

(1) Tetonic events recognized earlier for the European Prototethys and known for the North Atlantic belt are also clearly distinguishable in the evolution of the Urals–Okhotsk mobile belt. Therewith, the Cadomian tectogenesis was of less importance for the basins of the "northern" group of continents.

(2) Nearly synchronous events, including those diverse in action, in different belts and in different parts of a single belt appear to have been under dynamic conjugation. This is the most evident for the Vendian, Cadomian, and Salairian events.

(3) The Dalslandian tectogenesis, which occurred 1200(1100) - 1000 Ma ago, reflected one of the pronounced pulses in continental rifting that preceded the opening of ocean basins in all the mobile belts. During this period, compressive deformations, folding, and granitization in the mobile belts as well as the closing of aulacogens in the periphery of paleocratons were accompanied by intense tension and destruction of the continental crust.

(4) Phases of rifting prior to the opening of oceans appeared to be different. They were of short duration in the Mediterranean basin, especially in its peri-Gondwanian part. Oceanic sedimentation commenced here during the Late Riphean, and the opening of the Prototethys marked a separation between the "northern" and "southern" groups of continents. The Vendian rifting was of moderate duration and preceded the opening of the Iapetus in the North Atlantic belt and the appearance of major basins in the Central Asian ocean in the Urals–Okhotsk belt. These events accomplished the separation of the "northern" and "southern" groups of continents and initiated the breakup of the former.

(5) The asynchronous opening of ocean basins resulted in the time diversity of arc magmatism within their limits.

(6) Some of the principal tectonic events in the evolution of different mobile belts may be tentatively characterized as processes that compensated for one another. For instance, the epi-Cadomian orogeny in the closed West Asian segment of the Prototethys was compensated, to an extent, by the opening of the Central Asian ocean. The Kazakhstan–Tien Shan basin of this ocean appeared to be connected with the peri-Fennosarmatian relic basin in the northeast of the European Prototethys and prolonged its existence period to the Vendian–Cambrian boundary period. The crustal tension in the latter compensated for the collision in the peri-Gondwanian part of the Prototethys.

Against a background of dominantly closing basins in the Central Asian ocean during the Salair orogenic phase, the preservation of the Kazakhstan–Tien Shan segment was conjugated here with the reopening of the Uralian ocean, which may have been responsible for the closing of the peri-Fennosarmatian part of the Prototethys. In addition, the Salair tectogenesis in the peri-Fennosarmatian basin and the simultaneous closing of some marginal basins of the Iapetus were related to and

[2] The structural patterns of this basin were similar to those observed in the synchronous peri-Fennosarmatian part of the Prototethys (Rudakov, 1991).

partially compensated by the opening of the Rheicum ocean in the Central Europe and peri-Gondwanian basins in the Mediterranean belt, which were later integrated by the Paleotethys. In the Proterozoic-Paleozoic boundary period, the North Atlantic, Mediterranean, and Urals–Okhotsk mobile belts were interrelated as a single tectonic system of global rank.

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Reconstruction of Late Paleozoic Climatic Changes in Angaraland According to Phytogeographic Data

M. V. Durante

Institute of Geology, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Abstract – A brief description of the Carboniferous and Permian floras of the Angara (northern extratropical) paleofloral province is given. The latter includes lepidophyte flora from the Early Carboniferous, postlepidophyte flora from the beginning of the Middle Carboniferous, and cordaite flora from the second half of the Carboniferous–Permian, comprising three geofloras. Analysis of the compositions of floras and their geographical distributions allows the climates of the different Carboniferous and Permian epoches to be reconstructed: a warm and frost-free climate during the Early Carboniferous, a cold climate at the beginning of the Middle Carboniferous, and a temperate one during the second half of the Carboniferous and Permian. Abrupt cooling at the Early-Middle Carboniferous boundary (replacement of the frost-free climate by the cold one) and the synchronous Gondwanaland glaciation of the southern hemisphere suggest the onset of a contrasting climate and the formation of subpolar "refrigerators," which evidently existed by the end of the Late Permian. One epoch of relative cooling (the Early Permian–beginning of the Late Permian) is distinguished from the generally temperate climate of the second half of the Carboniferous–Permian. It separates two relatively warm epoches: the late Bashkirian–Late Carboniferous and the Late Permian epoches.

Key words: *Climate, Late Paleozoic flora, Angaraland, paleofloral, Subangara paleofloral region.*

It is well-known that the distribution of vegetation and landscapes on the Earth's surface depends mainly on the climate. Types of vegetation determine the character of landscape zones on the planet, which are usually reflected in their names.

This very fact allows phytogeographic investigations, including the history of the formation and dynamics of phytochories, to be considered as the main source of knowledge about climatic changes.

In additional to phytogeographic analysis, there exist a number of more traditional methods for defining the trend of climatic changes on the basis of paleobotanic data. The latter include analysis of the time-dependent increase or decrease in the diversity of floral assemblages, ascertainment of the dynamics of reconstructed plant formations, and verification of the diagnostic morpho-anatomical features of the plants themselves. All these methods were used to reconstruct the Late Paleozoic climatic changes in Angaraland.

A history of the Carboniferous and Permian floras of the Angara (northern extraequatorial) paleofloral region is given below. Floral evolution occurred against a background of increasing floral differentiation, growing differences in the floral compositions of near-equatorial (Euramerican and Cathaysian) paleofloral regions and Angaraland, and complication of the phytogeographic structure of Angaraland proper. A climatic interpretation of the paleofloral and phytogeographic changes is given.

THE EARLY CARBONIFEROUS: LEPIDOPHYTE FLORA

The formation of the Late Paleozoic phytogeographical situation started at the beginning of the Early Carboniferous, when the Angara continent was separated from both Euramerica and Cathaysia by vast marine basins. The Angara continent was occupied by impoverished and monotonous flora with a predominance of endemic lepidophytes with nondeciduous leaves (in contrast to the abundant and diverse floras of the other two continents).

During the first stage of the existence of lepidophytic flora (Tournesian), small-stemmed and small pillow forms with dichotomously ramified stems (endemic genera *Pseudolepidodendron* and *Ursodendron*, single species of *Tomiodendron*, and cosmopolitan *Eskdalia* genus) predominated. During the Visean and Serpukhovian, large treelike pillow forms with nonramified stems (*Tomiodendron* and *Angarophloios* genera), which were edificators of vegetation, became widespread. Another group of small-stemmed and herbaceous lepidophytes (*Caenodendron*, *Angarodendron*?, *Lophiodendron*, and *Pseudocyclostigma*) continued towards the lowlands and was less abundant than the first group. Primitive pteridosperms with cyclopteroid plumelets (*Abacanidium* genus), pra-ferns of the *Chacassopteris* genus, and rare horsetails were also restricted to lowlands (marine coasts and lakes).

Near the Early-Middle Carboniferous boundary, dominant forms of the lepidophyte flora – treelike

lepidophytes – became extinct. *Chacassopteris* and some horsetails (*Archaeocalamites*) also disappeared from the section. Meien (1968) was the first to note that mass extinction of the treelike lepidophytes happened in the middle part of the Ostrog Formation of Kuzbass. The fact that lepidophytes had timber of the manoxylic type, which is now characteristic of plants growing in areas with a frost-free subtropic climate, allowed him to correlate this event with the cooling. Apparently, climatic changes at this boundary (from a slightly differentiated frost-free climate to a climate with sharply isolated near-equatorial and extraequatorial areas) were of a global character.

BEGINNING OF THE MIDDLE CARBONIFEROUS: POST-LEPIDOPHYTE FLORA

At the beginning of the Middle Carboniferous, lepidophyte flora was replaced by extremely impoverished relic post-lepidophyte flora that was practically devoid of new forms. The latter probably corresponded to the maximum cooling. Pteridosperms of the *Abacanidium* genus were the main component of the post-lepidophyte flora. They produced an outburst of speciation and occupied the ecological niches previously covered by treelike lepidophytes.

Together with pteridosperms horsetail (*Paracalamites* and *Mesocalamites*), herbaceous (*Pseudocyclostigma*) and small-stemmed (*Angarodendron*? and *Caenodendron*) lepidophytes were widespread. As in the previous epoch, all of them were restricted to the lowlands.

Existence of the impoverished post-lepidophyte flora was short. Somewhere in the middle Bashkirian, it was displaced by relatively abundant and diverse pteridosperm-cordaite flora, which was the first among the three geofloras of the Late Paleozoic cordaite (polychronous) flora.

PROBLEMS OF THE ORIGIN OF THE ANGARA CORDAITES

Cordaites (mainly *Rufloria* with wide dorsal furrows) predominated in the pteridosperm-cordaite flora. Pteridosperms (*Angaropteridium*, *"Neuropteris," Cardioneura, Paragondwanidium*, and *Angaridium* genera) were also numerous. Spore plants played an important role only in certain parts of the sequence.

It should be noted that the pteridosperm-cordaite flora was represented mainly by new forms such as the Angara cordaites, which belonged to families and, probably, an order that were different from their Euramerican analogues. Apparently, these cordaites had no ancestors among the more ancient Angara floras, which were poor in seed plants. Thus, their origin is not clear.

We suggest Central Kazakhstan (the Kazakhstan province of the Euramerican region according to Meien, 1970) as the home of the Angara cordaites. Extremely impoverished archaeocalamite-lepidophyte-pteridosperm flora of the Euramerican type with abundant endemics existed in this region during the Early Carboniferous. In the Middle Carboniferous, it was replaced by essentially cordaite flora of the Angara type. This has usually been used to account for the migration of the Angara plants to the Central Kazakhstan area after the closing of the Ob'–Zaisan geosyncline (Meien, 1970).

Recent investigations have shown that the Angara cordaites in Central Kazakhstan were probably autochtonous. During the first half of the Middle Carboniferous (the Keregetas Horizon), their rare locations were confined to the Balkhash–Ili volcanogenic belt, while associations of the Euramerican type existed in the lowlands. As the Dzhungarskii marine basin and the lake–bog lowlands of northern Kazakhstan disappeared, the cordaite flora became more widespread. Plants were probably able to migrate to Angaraland proper during that time.

This migration was important in two respects. First, it gave rise to the polychronous cordaite flora of the Angaraland, which further evolved autochtonously. Second, it indicated the enlargement of Angaraland (owing to the inclusion of the Kazakhstan province) and the complication of its floral structure, because, despite what has been stated above, the flora of Central Kazakhstan differed significantly from that of Angaraland proper.

THE SUBANGARA PALEOFLORAL REGION

The floral isolation of Central Kazakhstan caused the formation of an outer (with respect to Angaraland proper) belt of peculiar floras, which Meien (1980) called the Subangara paleofloral region. In this case, the status of Angaraland rises to that of a paleofloral kingdom, subdivided into the Subangara and Siberian paleofloral regions.

The formation of the Subangara region (belt) was due to both floral and tectonic reasons, because it was mainly formed within microcontinents and continents that joined the Angaraland continent during the Late Paleozoic. The following main stages of this process are usually distinguished: (1) At the end of the Early Carboniferous, the Ob'–Zaisan geosyncline closed, which resulted in the unification of Angaraland and the Kazakhstan microcontinent. We consider the latter to include not only Central Kazakhstan, but also the eastern Urals and the northern Tien Shan. These areas were occupied by impoverished flora of the Euramerican type during the Early Carboniferous (Fig. 1). (2) During the middle of the Early Permian (Fig. 2), the Urals geosyncline closed, which resulted in the formation of the main part of the Subangara belt within the western Urals and the eastern Russian Plain. (3) At the end of the Early Permian (Fig. 3), floras of the Subangara type penetrated into northern Cathaysia.

The Subangara floras were more abundant and diverse and also more mixed regionally than the cor-

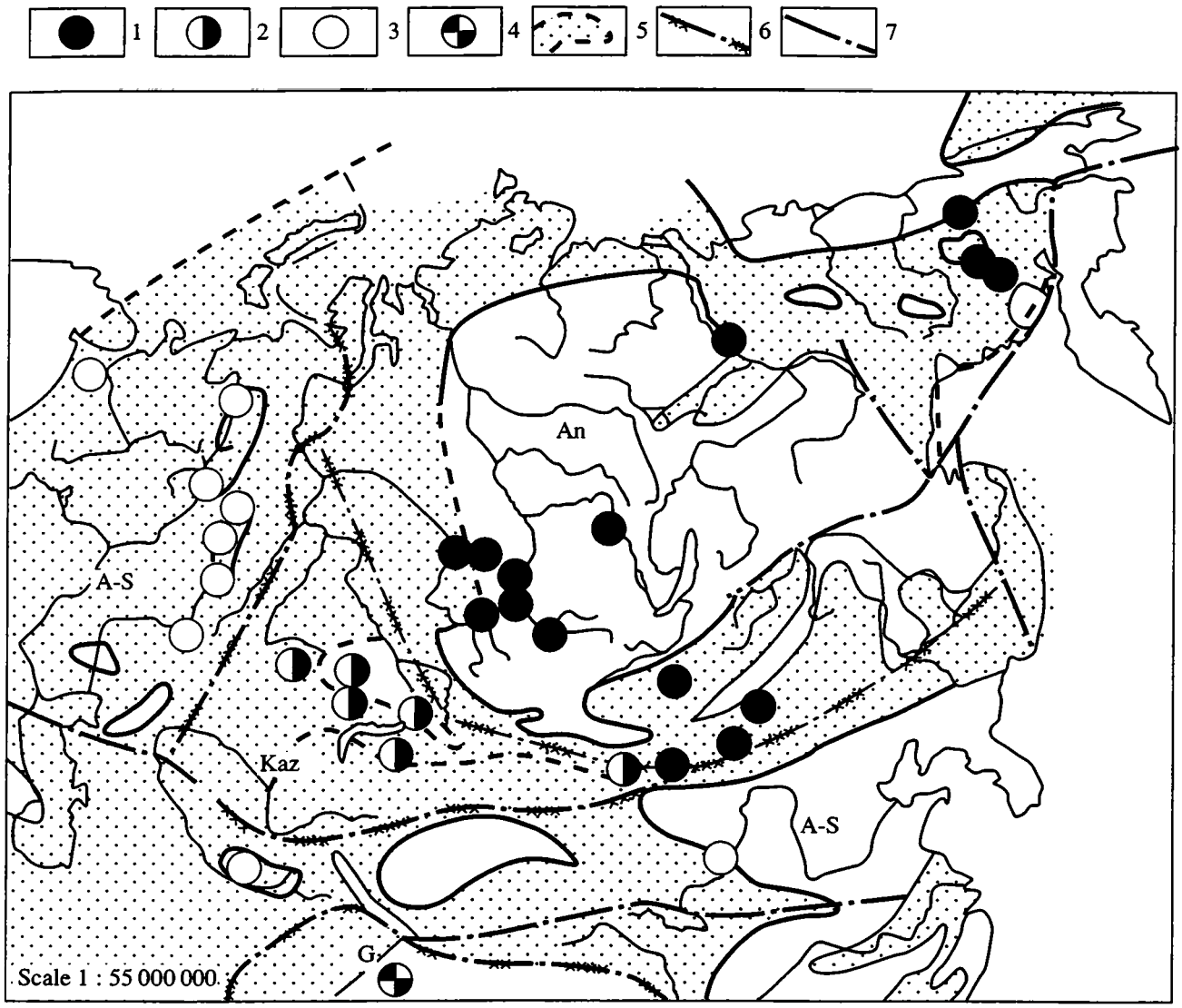

Fig. 1. Scheme of the paleofloral regions of northern and central Eurasia in the Early Carboniferous.
1 - 4 – Locations of plant remnants: 1 – the Angara paleofloral province (An), 2 – the Kazakhstan region (Kaz) of the Amerosinian paleofloral province, 3 – other regions of the Amerosinian paleofloral province (A-S), 4 – the Gondwanaland paleofloral province (G); 5 – marine basins; 6 – boundaries of phytochories of the first order: paleofloral provinces and kingdoms; 7 – the largest tectonic sutures.

daite flora of the Siberian region. Their originality increased upwards the section. However, evolution of the cordaite flora of the Siberian region occurred in close interaction with the Subangara floras. This interaction will be discussed below.

Coniferous plants unknown in the Siberian paleofloral region were numerous amongst Subangara floras of various ages. Permian floras of the Cis-Urals included abundant peltasperm pteridosperms, which rarely reached the central parts of Angaraland. They also included cardiolepids and various ginkgophytes. Spore plants were represented by endemic genera and species of horsetails and ferns uniformly distributed within the section, lepidophytes (*Viatcheslavia vorcutensis*) that were abundant in the lower part of the Permian sequence, and mosses that were found in the upper part of the sequence.

The presence of either typical Euramerican or typical Angara genera and species was also a characteristic feature of the Subangara floras. For example, *Calamites* of the Euramerican type were present in the Late Paleozoic floras of Central Kazakhstan, and typical Angara species of *Rufloria* were characteristic for the Early Permian of the Cis-Urals. The presence of taxa from regions adjacent to the Subangara paleofloral region allowed Meien (1980) to consider the latter as an ecotone region. This is only partly true, because during the greater part of the Late Paleozoic it was not only

a florogenic zone, but it also served as a barrier between the extraequatorial Siberian and near-equatorial Euramerican and Cathaysian paleofloral regions.

The endemism and originality of the Subangara floras grew upward the section. While the above-mentioned Middle-Late Carboniferous flora of Central Kazakhstan was similar to the pteridosperm-cordaite flora of Central Angaraland, the youngest Subangara flora (*Tatarina*) had nothing in common with synchronous Central Angara floras. It had no cordaites and was represented only by endemic genera: *Tatarina* (peltasperm), *Aequistomia* (cardiolepid), *Quadrocladus* (coniferous), and others.

As shown above, at the beginning of the Middle Carboniferous, the Subangara region was probably a maternal florogenic zone for the Siberian one, supporting it with the majority of the new forms of cordaite flora. However, the close interaction of the Subangara floras with the cordaite floras of Central Angaraland continued until the end of the Late Paleozoic.

THE SIBERIAN PALEOFLORAL REGION

The Late Paleozoic cordaite flora is the most completely displayed in the Siberian paleofloral region. It was less diverse than synchronous floras of Subangaraland, but its composition was more constant within the sequence and area. However, it is subdivided into three floras (geofloras) of different ages: pteridosperm-cordaite [second half of the Carboniferous–(probably) beginning of the Permian], mainly cordaite (Early–beginning of the Late Permian), and fern-pteridosperm-cordaite (Late Permian). Against a background of polychronous cordaite flora development, these three geofloras can be considered as corresponding to the phases of its origin, flourishing, and fading. Each phase was

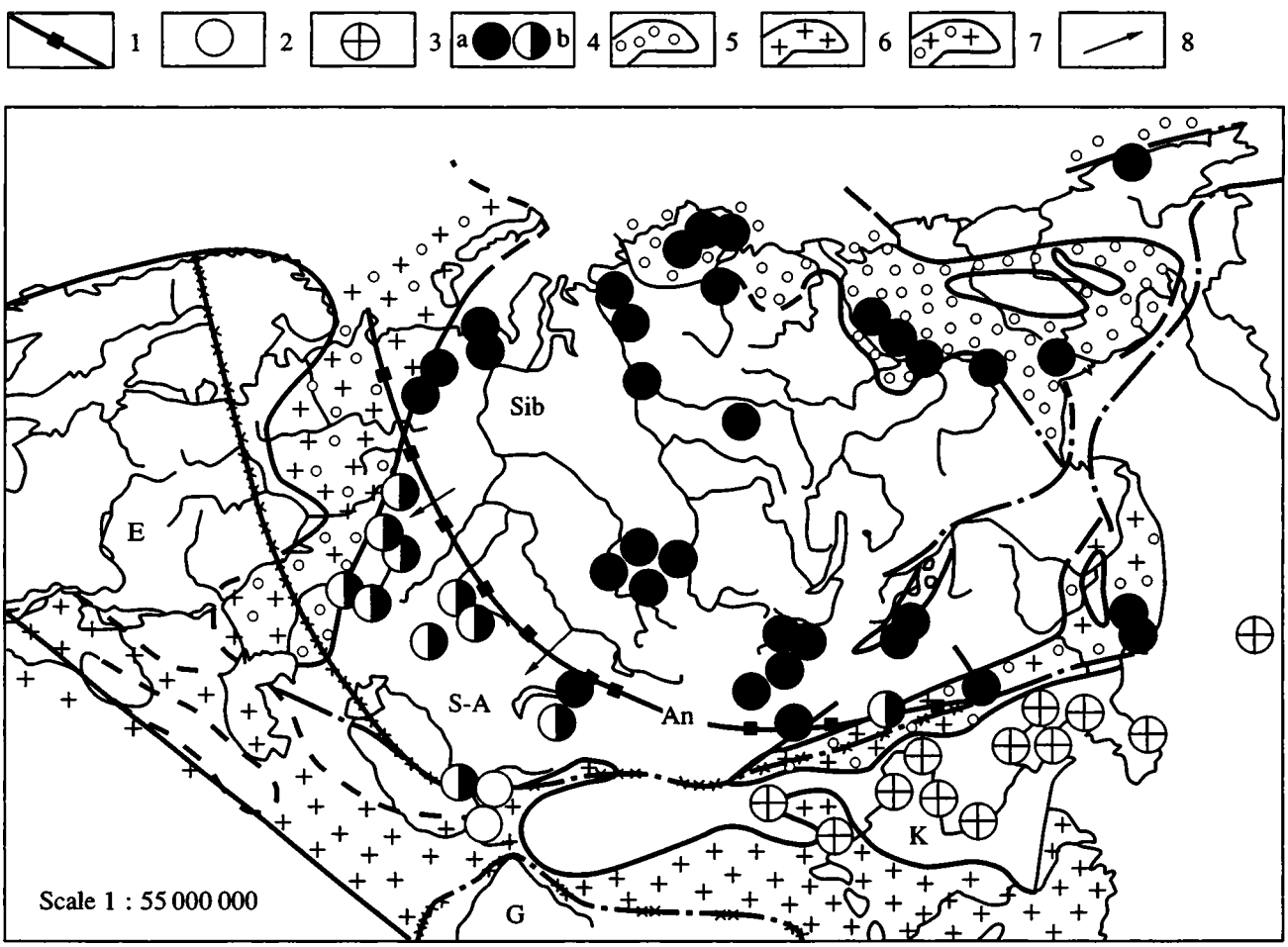

Fig. 2. Scheme of the paleofloral regions of the northern and central Eurasia in the Early Permian.

1 – Boundary between the Subangara (S-A) and Siberian (SiB) paleofloral provinces of the Angara paleofloral kingdom (An); 2 - 4 – locations of plant remnants: 2 – the Euramerican paleofloral province (E), 3 – the Cathyasian paleofloral province (K), 4 – the Angara paleofloral kingdom: a – the Siberian paleofloral province, b – the Subangara paleofloral province; 5 - 7 – marine basins with fauna of different paleobiogeographical types: 5 – boreal; 6 – Tethyan; 7 – mixed fauna; 8 – main direction of plant migration.

For other symbols, see Fig. 1.

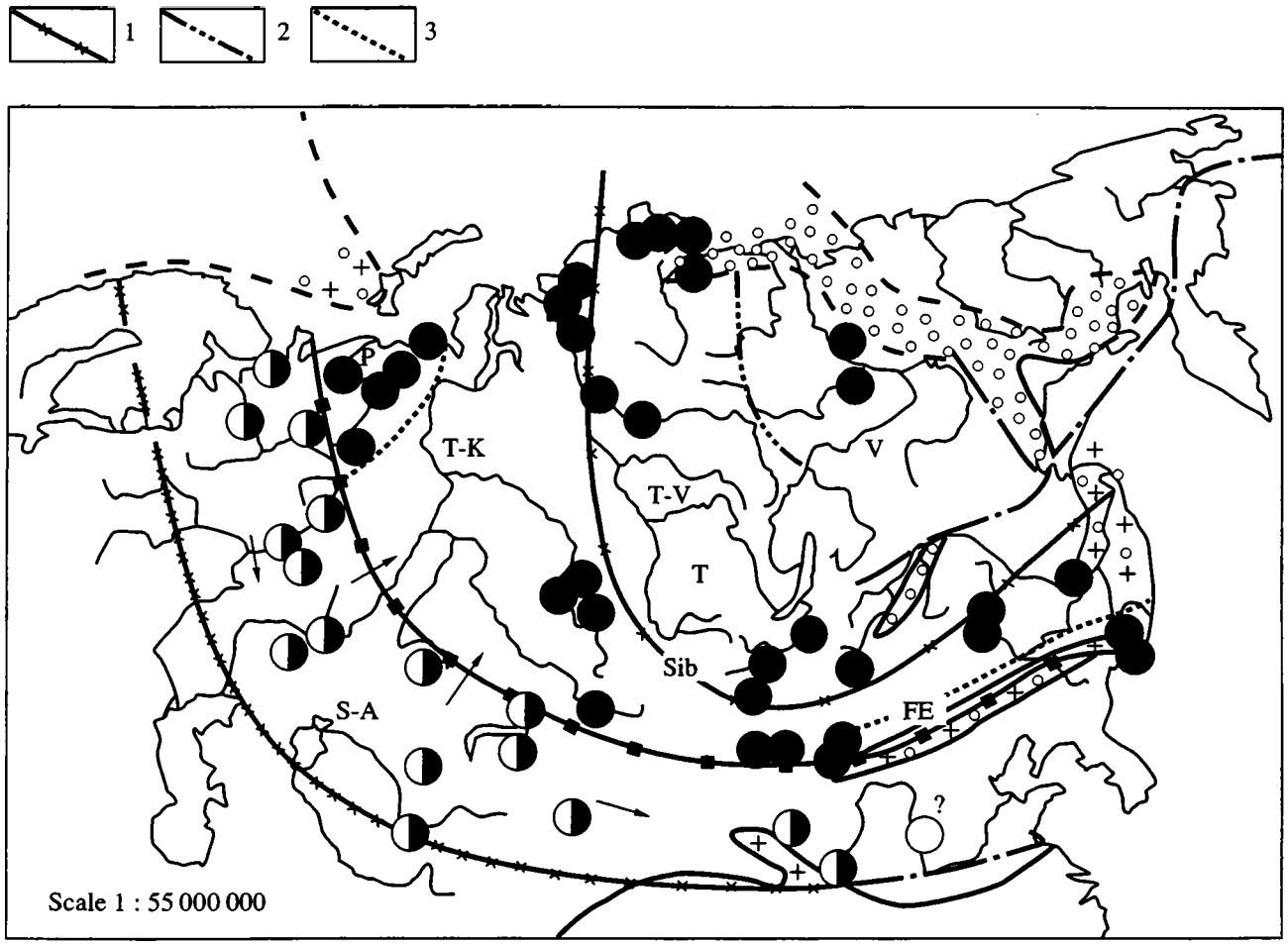

Fig. 3. Scheme of the paleofloral regions of northern and central Eurasia in the Late Permian.
1 - 3 – Boundaries of phytochories: 1 – boundary between the Taimyr–Kuznetsk (T-K) and Tunguska–Verkhoyansk (T-V) subprovinces of the Siberian paleofloral province; 2 – boundary of the Tunguska (T) and Verkhoyansk (V) paleofloral regions; 3 – boundaries of the paleofloral provinces: P – Pechora; FE – Far East.
For other symbols, see Figs. 1 and 2.

characterized not only by peculiar species composition, but also by a phytogeographical structure characteristic of the Siberian paleofloral region and by its own type of interaction with Subangaraland.

Thus, it is clear that the origin of the pteridosperm-cordaite, the oldest of the three cordaite floras, was mainly due to extensive migration from the Subangara paleofloral region. Cordaites (mainly *Rufloria* with wide dorsal furrows) predominating in it as well as some pteridosperms and horsetails (*Calamostachys* and small-leaved *Annularia* of the Euramerican type) were probably migrants. However, the pteridosperm-cordaite flora contained autochtonous elements that originated from the previous floras. The latter were represented by lepidophytes of the *Angarodendron* and *Caenodendron* genera; a predominating amount of the horsetails (*Paracalamites* genus); and pteridosperms with cyclopteroid and neuripteroid plumelets (*Angaropteridium* and *"Neuropteris"* genera), which were probably the descendants of *Abacanidium*.

The epoch of pteridosperm-cordaite flora existence was marked by the beginning of an internal phytogeographical differentiation in the Siberian region. The composition of floral assemblages throughout the entire paleofloral region did not vary significantly except in the central part: the Verkhoyansk subregion, including all of northeastern Russia from the eastern edge of the Siberian platform. In this region, together with *Rufloria* with wide dorsal furrows characteristic of the pteridosperm-cordaite flora, species of the same genus with narrow dorsal furrows of the "Permian" type were recorded.

The cordaite flora that replaced the pteridosperm-cordaite one was characterized by an overwhelming dominance of cordaites. Among other groups of vegetation, only horsetails (*Paracalamites, Phyllopitys,* and *Annularia*) played an important role in this flora. Pteri-

dosperms and ferns were represented by single genera and species. Leaves of cordaites not only predominated in floral assemblages, but they also reached their maximum size. Both *Rufloria* with narrow dorsal furrows and *"Cordaites"* without dorsal furrows were equally spread among them.

Floral composition was practically constant throughout the entire Siberian paleofloral region (except local variations). The essential feature of the cordaitic geoflora was that the cordaites characteristic for this geoflora actively penetrated into the Subangara paleofloral region. In other words, during the epoch of cordaite geoflora existence (Early - beginning of the Late Permian) the direction of plant migration was opposite to that of the previous epoch (Fig. 2).

The youngest Late Paleozoic geoflora of Angaraland – fern-pteridosperm-cordaite – was characterized by a significant increase in the systematic diversity of plants. Simultaneously, the role of cordaites and the size of their leaves decreased.

Ferns and endemic horsetails with vegetative sprouts belonging to *Phyllotheca* and *Annularia* genera were abundant among spore plants. Pteridosperms were extremely diverse. Two groups were identified among them: endemics (*Glottophyllum*, *Tychopteris*, and others) and migrants from the Subangara paleofloral region. The latter were mainly represented by callipterids and *Psygmophyllum*. Ginkgophytes (*Ginkgoites*) and cicadiphytes (*Tomia, Javorskya*, and *Guramsania*) appeared during the second half of the epoch of fern-pteridosperm-cordaite flora existence.

"Cordaites" without dorsal furrows predominated among primarily small-leaved cordaites. However, various *Rufloria*, which became extinct before the extinction of this flora, were widespread.

Upward the sequence, cordaiteless coniferous-fern (Korvunchana) flora of Triassic age replaced the fern-pteridosperm-cordaite flora.

One of the most essential features of the fern-pteridosperm-cordaite flora is its especially strong spatial differentiation (Fig. 3). For the period of the existence of this flora, Meien (Meien, 1970; Vakhrameev *et al.*, 1978) defined at least three subordinate zonal phytochories within the Siberian paleofloral region: the Taimyr–Kuznetsk subregion; the Tunguska–Verkhoyansk subregion; and, within the latter, the Tunguska and Verkhoyansk paleofloral areas. In addition, several azonal phytogeographical subdivisions are distinguished; the Pechora and Far East regions are the largest among them.

The strengthening of floral differentiation during the transition from the cordaite geoflora (the upper Balakhonka Subgroup of the Kuzbass) to the fern-pteridosperm-cordaite one (the Kol'chugino Group) correlates with another change in the direction of plant migration from the Subangara paleofloral region into the Siberian one. In fact, the boundary between the described geofloras coincides with the appearance of callipterids. As mentioned above, they may be considered as Subangara migrants. The distribution of callipterids is restricted to the Taimyr–Kuznetsk subregion. Later, the migration of *Psygmophyllum* and *Rhipidopsis* reached the peripheral parts of the Tunguska–Verkhoyansk subregion, but neither of them seemed to reach the Verkhoyansk area, where fern-pteridosperm-cordaite flora was represented by impoverished, almost completely cordaite associations.

CLIMATIC INTERPRETATION

The history of the Carboniferous and Permian floras of Angaraland and the study of the formation and dynamics of its phytogeographic structure allow rather definite conclusions to be made about the main tendencies of climatic changes. These tendencies are shown by the climatic curve in Fig. 4. More detailed climatic characteristics of the large Carboniferous and Permian epochs are given below.

(1) *The Early Carboniferous.* Meien (1968) was the first to show that the Early Carboniferous epoch with lepidophyte flora was characterized by a warm frost-free climate. This climate is indicated by the manoxylic-type stem structure, which is now characteristic of subtropical and tropical plants. Since Angaraland in the Early - Middle Carboniferous was situated in low and middle latitudes (Gorodnitskii *et al.*, 1978; Zonenshain *et al.*, 1987; Khramov *et al.*, 1982), its frost-free climate probably indicates that the climate of the northern hemisphere and the entire Earth was uniform and relatively warm.

Lithologic characteristics of the Lower Carboniferous deposits of Angaraland (the presence of bauxites and variegated sediments) confirm these climatic reconstructions.

(2) *Climatic change during the Middle Carboniferous.* As was shown above, the mass extinction of tree-like lepidophytes near the Early-Middle Carboniferous boundary indicates a climatic change in the Middle Carboniferous. Meien's (1968) assumption that the extinction of lepidophytes was caused by cooling is supported by both the thermophilic nature of the tree-like lepidophytes and the relic character of the extremely impoverished flora, virtually devoid of new forms, that replaced the lepidophyte. This assumption is also supported by the forestless "tundra" vegetation of the post-lepidophyte time that replaced the peculiar lepidophyte "forests" and existed in Angaraland until the middle Bashkirian.

Similar climatic changes occurred in Gondwanaland (*The Carboniferous* ..., 1985). At the same time, rich and diverse flora (Wagner, 1984) with corresponding diverse forest vegetation (without sharp changes) evolved virtually without interruptions in near-equatorial regions during the entire Carboniferous. This character of Carboniferous flora and vegetation indicates an absence of catastrophic temperature oscillations in

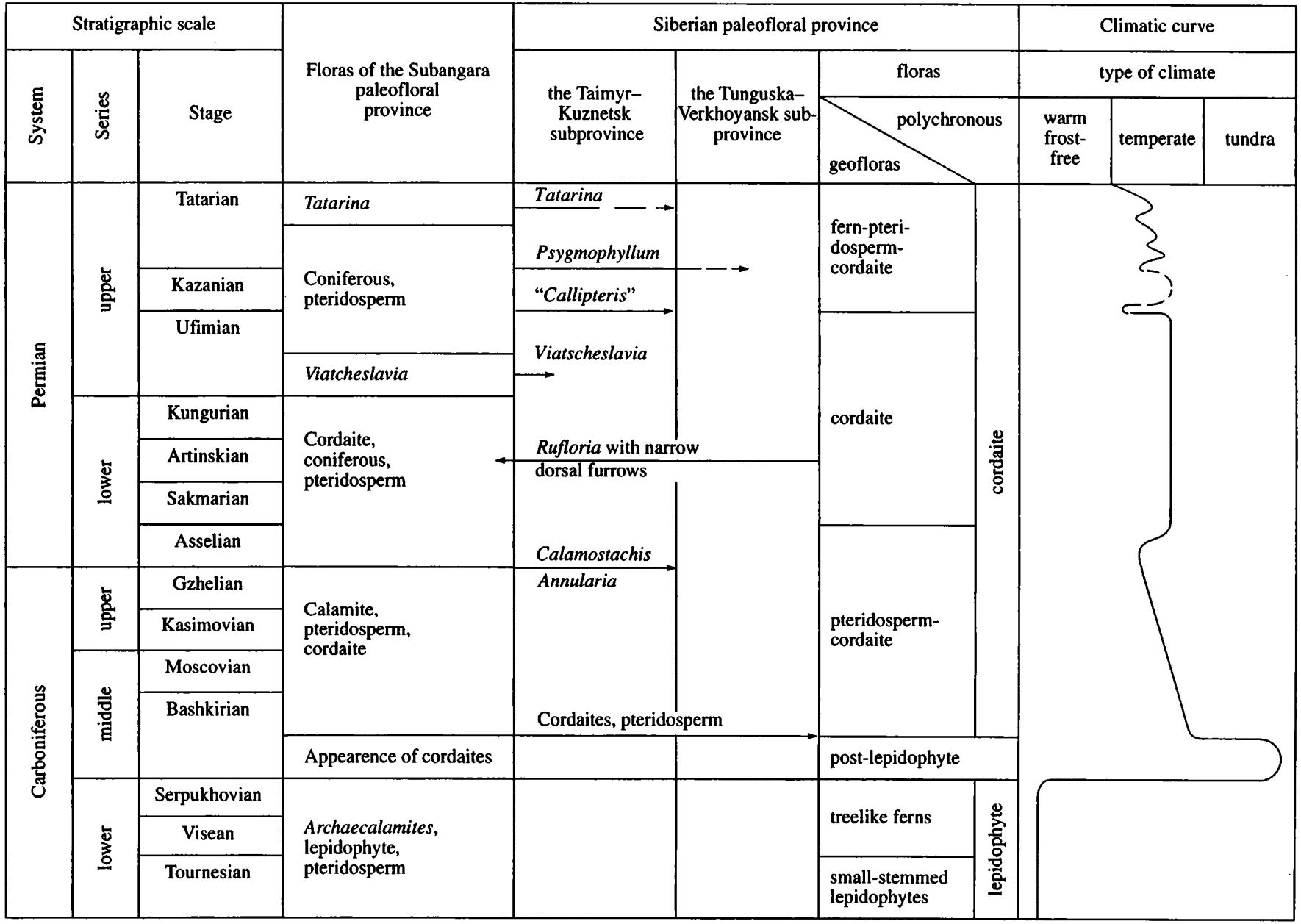

Fig. 4. Interactions of floras of the peripheral (the Subangara paleofloral province) and central (the Siberian paleofloral province) parts of the Angara kingdom. The arrows mark the direction of plant migrations.

near-equatorial areas. Thus, global cooling during the Middle Carboniferous was manifested only in extratropical regions and was related to the formation of subpolar "refrigerators."

These conclusions agree with the first findings of Late Paleozoic tillites, indicating the beginning of Late Paleozoic glaciation in the southern hemisphere, near the Early - Middle Carboniferous boundary (*The Carboniferous ...*, 1985). It is assumed that Middle Carboniferous glacial-marine deposits are present in the northern hemisphere (Ustritskii and Yavshits, 1971).

(3) *The Late Paleozoic climate of Angaraland.* Despite the fact that the Carboniferous cooling of Angaraland was short and followed by a relative warming during the middle Baskirian, which resulted in cordaite flora spreading, the warm and frost-free climate of the Early Carboniferous was not restored. Sharp climatic differences between tropical and extratropical regions were characteristic of the entire Late Paleozoic. The complete absence of plants with manoxylic stems and, conversely, the predominance of cordaites with picnoxylic stems with clear growing rings marking relatively long unfavorable seasons show that the climate of Angaraland at that time was colder than in the Early Carboniferous. Floral (sharp differences between tropical and extratropical floras and the absence of cosmopolitan elements in the latter) and phytogeographical data also indicate climatic contrasts.

Compared with the previous (Fig. 1) and subsequent (Dobruskina, 1982) epochs, the phytogeographical situation in Angaraland was extremely complicated (Figs. 2 and 3). Its outer boundary became almost impenetrable because of the Late Paleozoic Subangara "ecotone" belt at the margin of Anagaraland. As mentioned above, this belt played an important role in the formation and evolution of cordaite flora, and it also served as a barrier that prevented direct exchanges between Angara and neighboring near-equatorial floras.

It should be stressed that geographical isolation was not the cause of the highly endemic Late Paleozoic flora of Angaraland. Actually, during the Carboniferous and Permian this isolation gradually decreased (Figs. 1 - 3), while, conversely, phytogeographical isolation increased. Apparently, only the high degree of Late Paleozoic climatic contrasts can be suggested as an explanation for this phenomenon.

In turn, climatic contrasts probably existed due to the stable subpolar "refrigerators" that formed during the Carboniferous cooling, which maintained the high-temperature gradient between near-equatorial and extraequatorial regions during the entire Late Paleozoic. Available data concerning the occurrence of glacial-marine deposits at several levels in the Upper Paleozoic sequence of the northeastern Russian Plain (Andrianov, 1966; Ustritskii and Yavshits, 1971; Epshtein, 1972; *Opornyi Razrez ...*, 1990) suggest the existence of a cold marine basin near the northern pole. Unlike the southern hemisphere, either there were no glacial caps in the northern hemisphere or they existed locally for a short time. Though reliable data on Late Paleozoic temperatures in Anagraland are now lacking (Yasamanov, 1985), the climate of the majority of this period (except the very beginning of the Middle Carboniferous) may be regarded as a temperate humid one. This is supported, first, by a comparison with the modern situation. The seasonal climate of the middle latitudes ensured the stable existence of forest vegetation ("cordaite taiga") for a long time. It was probably similar to the present temperate climate. Secondly, the same inference was made after a comparison of this climate with the warmer one of the Early Carboniferous and the more severe climate of the maximum Late Paleozoic cooling, which occurred at the beginning of the Middle Carboniferous. The warm frost-free climate of the Early Carboniferous, without distinct seasons, was probably similar to the modern subtropical climate, and the absence of forest vegetation and the spreading of "tundra" landscapes indicate a cold climate at the beginning of the Middle Carboniferous. Thus, the intermediate climate of the greater part of the Late Paleozoic may be regarded as a temperate one.

Extensive coal accumulation in Anagraland during the Late Paleozoic proves the climate to be humid.

Our conclusions about a temperate Late Paleozoic climate in the middle latitudes do not concern the entire Angara kingdom, but only its largest central part: the Siberian paleofloral region, which was covered with cordaite flora. As stated above, the Late Paleozoic floras of the peripheral Subangara region were characterized by a significant admixture of Euramerican (tropical and subtropical) elements and were richer and more diverse. Thus abundance and diversity of floras together with the absence of coal accumulation and the spreading of red beds indicate a hot and dry climate. Sometimes it seems to have been arid (the appearance of salt-bearing and gypsum-bearing horizons in the sequence).

The climate of the Subangara paleofloral region is not the subject of this article. Interrelations between the Subangara and Siberian floras, involving the ascertainment of the migration routes of plants from the periphery of Angarida to its central parts or vice versa, are especially important for the reconstruction of the largest climatic oscillations during the Late Paleozoic.

As was shown above, each of the three epochs that correspond to the three geofloras comprising polychronous cordaite flora is characterized by its own direction of plant migration within the Angara kingdom. Reconstructions of the Late Paleozoic climate, shown by the climatic curve (Fig. 4), are based on the study of plant migrations, changes in the composition of the cordaite assemblages, and peculiarities of the phytogeographic differentiation.

Climatic characteristics of the second half of the Carboniferous (pteridosperm-cordaite geoflora), the Early-beginning of the Late Permian (cordaite flora), and the Late Permian (fern-pteridosperm-cordaite geoflora) are given below.

1. MIDDLE-LATE CARBONIFEROUS

The described floral data do not confirm the supposition that the Earth's climate during the second half of the Carboniferous was characterized by a progressive cooling that reached its maximum by the beginning of the Permian (Yasamanov, 1985). The maximum cooling of Angaraland, which is marked by the spreading of forest-free "tundra" landscapes and, probably, a cold climate, coincides with the beginning of the Middle Carboniferous. However, this cold climatic epoch is restricted to the first half of the Bashkirian. In the second half of the Bashkirian, a temperate climate replaced the cold one. This is marked by the spreading of cordaite flora and the appearance of coal layers. As stated above, the cordaites probably migrated from the Kazakhstan region. Their spreading within Angaraland (the Siberian paleofloral region) indicates a grandiose floral expansion that emphasizes the significance of the Bashkirian warming.

This expansion also indicates a relative smoothing of the climatic contrasts between the Siberian and the Kazakhstan paleofloral regions. During the Early Carboniferous, the latter was confined to the periphery of the Euramerican paleofloral region; however, in the Late Paleozoic it became a member of the peripheral Subangara belt of Angaraland. The reorientation of the Kazakhstan region and the "angarization" of its flora occurred during the second half of the Carboniferous. However, climatic differences between the Kazakhstan region and Angaraland itself still remained. The process of "angarization" during the Middle-Late Carboniferous Kazakhstan floras correlates with the termination of coal accumulation, i.e., it probably denotes aridization of the climate. An opposite correlation is observed in the Siberian paleofloral region: the spreading of cordaite flora coincides with the initiation of coal accumulation. Thus, during the second half of the Carboniferous, the boundary between the Siberian paleofloral region and the Kazakhstan region (the Subangara paleofloral region) coincided with the border between the temperate humid and the temperate but more arid climate.

Other (temperature) parameters seem to determine the isolation of the Verkhoyansk subprovince within central Angaraland. In this region (northeastern Russia), pteridosperm-cordaite geoflora existed under temperatures slightly lower than those in the other parts of Angaraland, which is indicated by the abundance of peculiar large cordaites typical of younger cordaite geoflora of other regions. The latter, as shown below, existed under cooler climatic conditions, which is indicated by the occurrence of glacial-marine sediments (Ustritskii and Yavshits, 1971).

In general, a clearly expressed warming was the main climatic feature of the second half of the Carboniferous in most of Angaraland. This did not involve the middle Bashkirian warming alone. An increase in the diversity of pteridosperm-cordaite flora upward the sequence and the appearance of several purely Euramerican taxa in Middle Siberia during the last stage of the existence of this flora show that the warming trend continued up to the boundary between the pteridosperm-cordaite and the younger cordaite floras.

2. THE EARLY PERMIAN–BEGINNING OF THE LATE PERMIAN

Until recently, it was thought that the boundary between the pteridosperm-cordaite and cordaite geofloras corresponding to the boundary between the lower and upper Balakhonka subgroups of the stratotype section of the Kuznetsk basin coincided with the Carboniferous–Permian boundary. Recent data suggest that the boundary between these systems should be placed within the pteridosperm-cordaite geoflora. This implies that the climate at the very beginning of the Permian probably did not differ from the climate described above. However, it was rather quickly replaced by a cooler climate corresponding to the cordaite geoflora.

Significant cooling at the boundary between the pterisosperm-cordaite and cordaite geofloras is proved, first, by the decreased diversity of the cordaite geoflora, which consisted almost entirely of cordaites. Secondly, the change in the direction of plant migration occurs at the boundary between the geofloras. The upper Balakhonka cordaites actively penetrated into the Subangara belt. By the middle of the Artinskian, the latter had formed as a large arch stretching from the northern Tien Shan to the Cis-Urals (Fig. 2). Expansion of the Angara cordaites was recorded in different parts of the belt: the northern Tien Shan (Kuraminskii Ridge), central Kazakhstan (Kyzyl-Kiya Formation of the Bakanasskaya depression and its analogues), and the Cis-Urals (the upper Artinskian and Kungurian deposits of numerous locations). The fact that this expansion occurred everywhere can only be explained by climatic changes.

It should be noted that the composition and characteristics of the cordaite geoflora throughout the Siberian paleofloral region are extremely uniform. This uniformity probably demonstrates the uniformity of climatic conditions throughout this vast region over a long period of time (the entire Early Permian and, probably, the Ufimian).

3. THE LATE PERMIAN (THE KAZANIAN AND TATARIAN STAGES, APPROXIMATELY)

As shown above, the youngest of the three geofloras of the Late Paleozoic cordaite flora – fern-pteridosperm-cordaite – differs from the previous one. Its

taxonomic composition is more diverse. The direction of the main routes of plant migration (from the Subangara to the Siberian region) changed. Floral differentiation, which obviously reflected the differentiation of the Late Permian climate of Angara, sharply increased.

Differentiation of the Late Permian climate was stressed by the behavior of the Subangara migrants. Most of them (callipterids, *Psygmophyllum*) penetrated only to the peripheral part of the Siberian region, forming a zone of relatively abundant and diverse floras: the Taimyr–Kuznetsk subregion. Only a few migrants (*Rhipidopsis*) are known within the Tunguska syncline (the Tunguska area of the Tunguska–Verkhoyansk subregion). None were found in the central (the Verkhoyansk) region of the same subprovince, which consists of impoverished floras. The observed migrations show that the Late Permian warming was characteristic only of the peripheral part of the Siberian region; the climate remained cool in the central parts.

The climate of the Late Permian epoch considerably varied over time. This is displayed by the contrasting Upper Permian sequence of the Omolon massif, which is now one of the Late Permian territories available for study in high latitudes. The lower and upper sequences of this section (the Omolon and Khivach horizons) consist of mainly calcareous deposits; the middle sequence (the Gizhiga Horizon) is characterized by abundant diamictites, which are usually interpreted as glacial-marine sediments (*Opornyi Razrez ...*, 1990). Findings of diamictites not only confirm the existence of the subpolar Late Permian "refrigerators," but also account for several peculiarities in the migrations of plants, in particular, the uneven distribution of migration waves and their different directions.

Migrations from the margins of Angaraland to the central parts, in association with warmings, are restricted to the beginning (callipterids) and the second half of the epoch of fern-pteridosperm-cordaite flora existence (*Psygmophyllum*, *Teniopteris*, and *Rhipidopsis*). The Subangara migrants are absent in the middle part of the Upper Permian sequence of Siberia (nearly the entire Il'inka Subgroup of the Kuznetsk basin and its analogues), which is correlated with the Gizhiga Horizon. Conversely, an opposite migration was probable, because *Rufloria* of the "Il'inka" type were found in several locations of the Subangara belt. The destruction of the Late Permian "refrigerator" in central Angaraland was probably caused by progressive warming at the Permian–Triassic boundary. This is proved by the fact that Triassic floras are less differentiated than the Late Permian ones.

CONCLUSION

Here, we provide a short summary of the dynamics of the phytogeographical situation and climate of the northern extratropical (the Angara) province (kingdom) of paleoflora during the Carboniferous and Permian.

(1) As shown above, the climate of the northern extratropical area during the Carboniferous and Permian varied considerably (Fig. 4) from nearly subtropical in the Early Carboniferous to "subarctic" (forestless "tundra") at the beginning of the Middle Carboniferous. A temperate humid climate prevailed during the greater part of the Late Paleozoic (except the mentioned epoch of "subarctic" climate).

(2) A distinctive feature of the Late Paleozoic phytogeographic and climatic situation is a more clearly displayed spatial differentiation of floras than that which was observed during the previous (Early Carboniferous) and subsequent (Mesozoic) epoches. This differentiation indicates contrasting climates during the Late Paleozoic climate. Sharp differences between the extratropical cordaite flora of Angaraland and the richer and more diverse tropical floras as well as significant internal differentiation of the floras of the Angara and Amerosinian paleofloral kingdoms are the main indicators of climatic contrasts.

Climatic contrasts confirm the supposition (Chumakov, 1978, 1984) that the entire Late Paleozoic belongs to the glacial age, because these contrasts indicate the existence of subpolar "refrigerators."

(3) Analysis of the Late Paleozoic phytogeographical situation allows us to make some conclusions about the climatic zonality of the cold periods (glacial eras) as opposed to that of the warm periods (thermal eras, according to Chumakov, 1978, 1984). It is generally accepted (Markov *et al.*, 1978) that, compared to warm periods, glacial ages are characterized by more intensive climatic differentiation and significant migration of climatic zones toward the equator. However, the paleofloral data are not indicative of these significant migrations of phytogeographic and, correspondingly, climatic boundaries.

The boundary of highest rank between the low latitudinal (near-equatorial) floras and those of temperate and high latitudes (extratropical) probably existed throughout the entire Phanerophyte except the Devonian (Vakhrameev *et al.*, 1978). During the Late Paleozoic, it was the outer boundary of Angaraland. After tracing the migrations of this boundary during the Phanerophyte, Meien (1969) came to the conclusion that it was rather stable over time. Of course, during different geologic periods, this boundary separated areas with different types of climates, thus proving the fact that secular climatic variations were reflected in more than just latitudinal migration of certain climatic zones.

(4) Climatic contrasts are reflected by a decrease in the penetrability of the boundaries between tropical and extratropical floras, rather than by the migration of paleofloral boundaries toward the equator. In the Late Paleozoic, interchange of taxa between Angaraland proper (the Siberian paleofloral region) and the Amerosinian paleofloral kingdom was completely absent due to the formation of the Subangara "ecotone"

belt. However, in the previous (Devonian–Early Carboniferous) and subsequent (Mesozoic) epochs such boundaries were not an obstacle for taxonomic interchanges.

In the Late Paleozoic, all of Angaraland existed as a closed system, in which the described correlation between temperature and the extent of floral differentiation was not sustained. For example, cooling at the Early-beginning of the Late Permian caused smoothing rather than growth of floral (climatic) differentiation in Angaraland (Fig. 2), while the Late Permian warming resulted in a growth of differentiation (Fig. 3). As was shown (Markov *et al.*, 1978), temperature oscillations usually have a different result.

In general, the Late Paleozoic history of Angaraland shows that the complex of phytogeographical units probably has a system character. Apparently, this is the reason for the stability of many phytogeographical boundaries over time. The fact that similar climatic changes caused different paleofloral effects is a result of this system character.

The described system of phytochories of Angaraland, including the Subangara belt and less stable phytogeographical units, was characteristic for the Late Paleozoic glacial era only. In the previous (Devonian–Early Carboniferous) and subsequent (Triassic) epochs (Vakhrameev *et al.*, 1978; Dobruskina, 1982), the phytogeographical structure of the Eurasia was different. Apparently, large phytogeographical structures correspond to major climatic eras. However, biotic factors are undoubtedly quite important to the evolution and functioning of such systems. The goal of our future investigations is to separate the biotic and climatic factors of the system.

To reconstruct the Late Paleozoic climate of Angaraland and compile a climatic curve, we used mainly data on plant migrations. Some other factors of a climatic nature may be identified in the future when the regularities of the functioning of phytochones is better clarified. This will provide more details of the curve. At present, our climatic curve should be regarded as a preliminary one.

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Endemic and Cosmopolitan Assemblages of Foraminifers and Ostracods from the Middle Jurassic Basins of Syria

K. I. Kuznetsova and M. R. Dobrova

Institute of Geology, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Abstract – Studied foraminifers and ostracods from Syria were recognized as composing endemic, cosmopolitan, and mixed assemblages that populated the Middle Jurassic basins of the eastern Mediterranean. The defined orictocoenosis composition was dependent on the bathymetric zonation of a basin. In various parts of the studied region (the Anti-Lebanon, the Coastal Mountains, the Kurd-Dag, and the Palmyrides), coeval assemblages of foraminifers and ostracods differed greatly in their taxonomic compositions and could be assigned to different fauna groupings or even fauna types. It was determined that the endemic Tethyan assemblages populated the shallowest basin parts (littoral and sublittoral zones), whereas other groups dominated by cosmopolitan forms inhabited deeper areas (lower sublittoral and epibathyal zones). Studied facial sequences of deep to shallow coastal Jurassic sediments from basins of the tropical climatic zone showed that foraminifer and ostracod dispersals were basically dependent on depth and related temperature factors, because the benthic organisms of both groups reacted primarily to changes in the water thermal regime near the bottom. The salinity factor mainly affected the distribution of ostracods, which were subdivided into assemblages dominated either by eurihaline forms or by species adaptable to the desalination of the basin waters.

Key words: *Middle-Upper Jurassic, endemic/cosmopolitan assemblages, facies successions, bathymetric/thermal factors.*

Studied thanatocoenoses of the Jurassic foraminifers and ostracods collected by geological investigations in Syria during the period 1986 - 1990 revealed many important features of their stratigraphic, areal, and facial distributions in this region. Time-related changes in these assemblages were distinguished after a bed-by-bed investigation of sections and were used to identify various stratigraphic units (zones, ecozones, and fauna-bearing beds) and to add necessary details (Fig. 1) to the Jurassic stratigraphic scale of Syria (Kuznetsova *et al.*, 1991; Kuznetsova, 1994; Kuznetsova and Dobrova, 1994). However, we encountered certain problems in correlating the Jurassic deposits of this region on the basis of these faunal groups. The procedure was especially troublesome when principal orogenic structures of Syria with abundant Jurassic rocks, i.e., the Anti-Lebanon, Coastal Mountains, Kurd-Dag, and Palmyrides (Fig. 2), were under comparison. Inside each of these structural regions, the Jurassic sections can be well-correlated without serious problems. It is remarkable that foraminifer and some ostracod assemblages from these regions differ not only in their species and genera composition, but also at the family level, thus representing different fauna types or groupings (Gordon, 1970; Basov, 1974, 1991; Basov *et al.*, 1972; Kuznetsova, 1994). These features are responsible for the correlation problems mentioned above, although the age intervals of different assemblages are well-constrained by their stratigraphic position in sequences; by other coexisting faunas (brachiopods, bivalves, and corals); and by cooccurring species of foraminifers and ostracods known from corresponding deposits of western Europe, Africa, the Caucasus, the Crimea, Madagascar, Canada, and other regions.

By analyzing the taxonomic composition of benthic foraminiferal assemblages that inhabited the Jurassic seas of the eastern Mediterranean, we distinguished the three following groups among them: (1) tropical Tethyan endemics from the equatorial areas of the Tethys, (2) sub-Tethyan forms with a wider distribution, and (3) cosmopolitan boreal elements that are very widespread but have narrow stratigraphic ranges of occurrence. The definitions of the first and second groups are well-known; however, the third one requires additional explanation. As follows from the name, this group occurs everywhere, including the Boreal provinces. Representatives occurring in marine Jurassic deposits of higher latitudes are very diverse genera and species of the families Ichthyolariidae, Nodosariidae, Lenticulinidae, Vaginulinidae, Polymorphinidae, Ceratobuliminidae, and Epistominidae, which define the nodosariid-epistominid type of this fauna (Basov, 1974). High evolutionary rate at the species level, associated with long existence periods of certain genera, is another characteristic feature of this group. For example, rapidly evolving species typify such foraminifer genera as *Lenticulina, Epistomina, Lamarckella,* and *Lamarckina.*

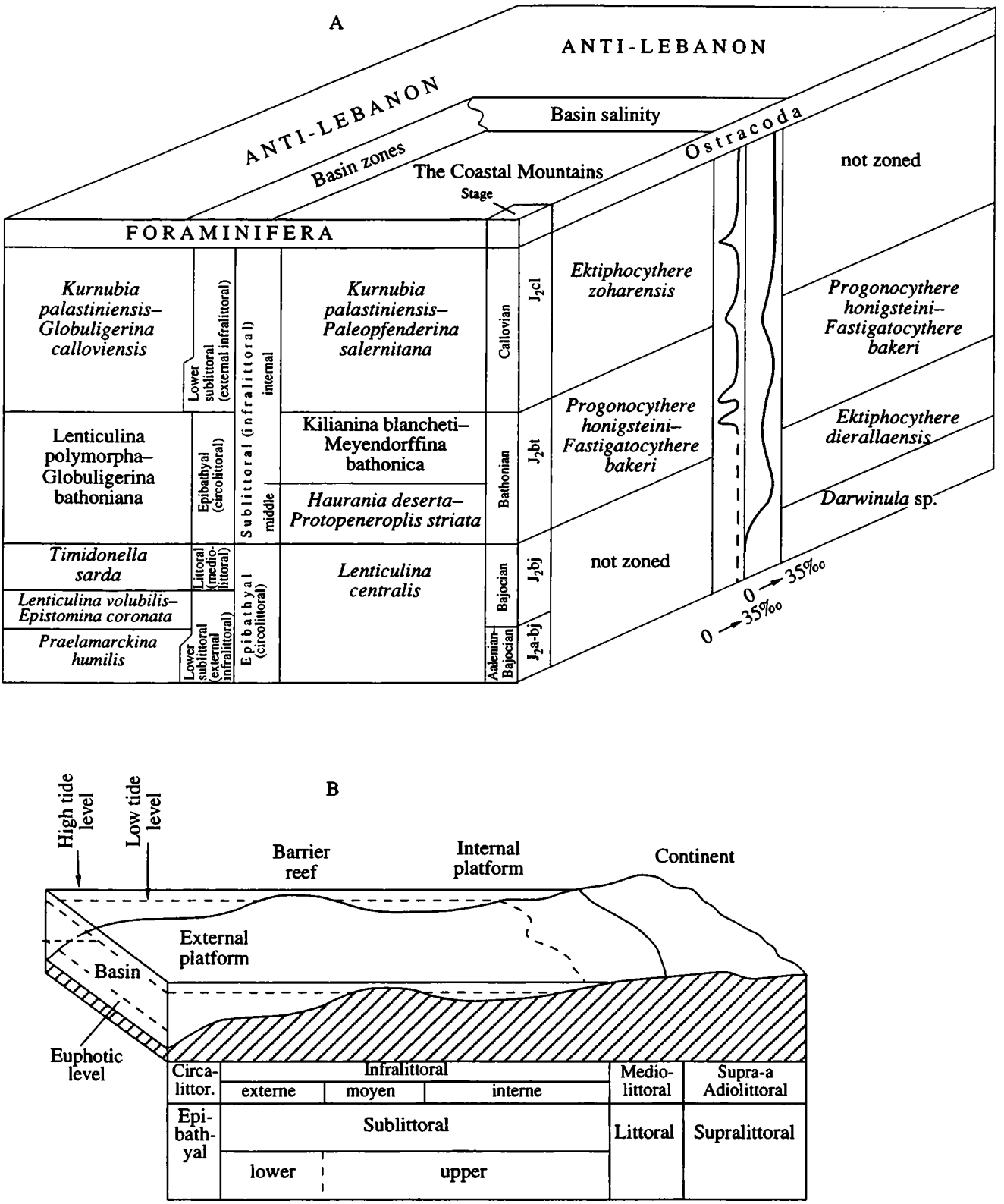


Fig. 1. Correlative foraminifer and ostracod zonation scales and bathymetric zones of corresponding salinity in the Middle Jurassic basins of Syria (A); model basins of the carbonate platform (B), modified after Pelissie *et al.* (1984).

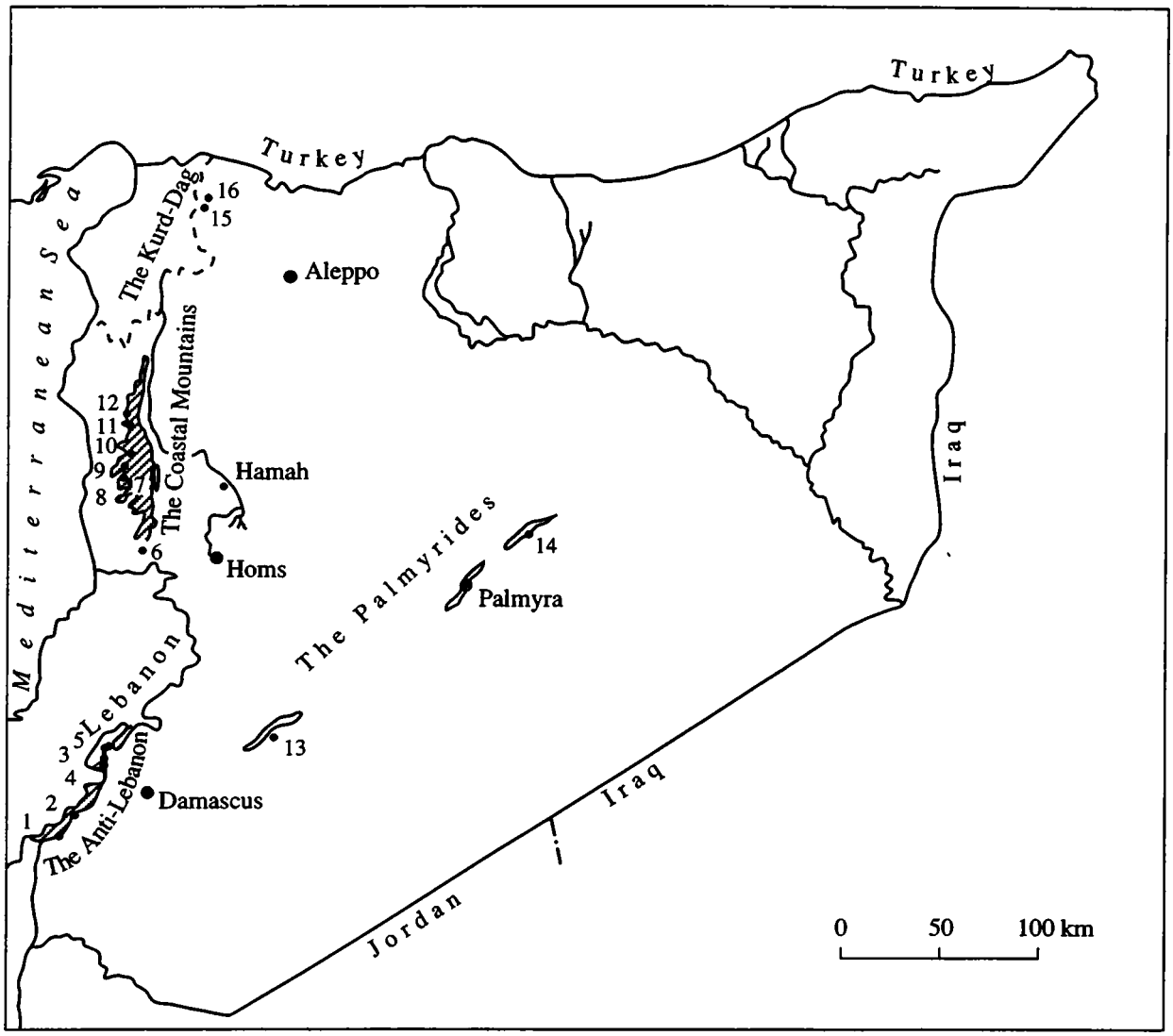


Fig. 2. Localities of studied sections and outcrops of the Jurassic marine deposits in Syria.
The Anti-Lebanon: 1 – Hadar, 2 – Arneh, 3 – Wadi Al Karn, 4 – Sad Al Karn, and 5 – Rowda. The Coastal Mountains: 6 – Nicola Quarry, 7 – Qadmous, 8 – Wadi Jannam, 9 – Jdaida, 10 – Mareen the East, 11 – Bab Janneh, and 12 – Chuekha. The Palmyrides: 13 – Al Zbeidi and 14 – As-Sattikh. The Kurd-Dag: 15 – Rajo and 16 – Smalek.

In the habitat basins, these taxa were usually dispersed in deeper zones than the Tethyan endemics, e.g., lituolids with large faveolate or alveolar shells, pfenderinids with a complex inner skeleton, dictyoconids, and other taxa.

Planktonic foraminifers were rare in all three Jurassic groups. They were least abundant among the cosmopolitan boreal fauna; in this case, they occurred only at stratigraphic levels corresponding to periods of transgressions: the Bathonian, Callovian, and Oxfordian.

In our work, we use the classification of basin zones adopted in Russia (Gorbachik *et al.*, 1970) along with the terminology of foreign researchers (Pelissie *et al.*, 1984). In the illustrative block-diagram (Fig. 1) featuring a basin in a carbonate platform, which was drawn using data on southwestern France (Pelissie *et al.*, 1984; Fig. 3), this terminology is given.

The ostracod fauna from the Jurassic deposits of Syria can be subdivided into four groups: (1) genera occurring only in the southern peri-Tethyan regions, (2) genera distributed in both the southern and northern regions fringing the Tethys paleoocean, (3) cosmopolitan taxa, and (4) genera with a wide geographic distribution that originated under peculiar desalination environments.

The first group includes the poorest fauna composed by genera *Glyptogatocythere, Zergacythere,* and *Virgulacytheridea*. The second one is the richest in taxa, which inhabited the southern and northern boundary basins of the Tethys (*Macrodentina, Procytheridea,*

Progonocythere, Fabanella, Oligocythereis, Cytherelloidea, Ektyphocythere, Glyptocythere, Fastigatocythere, Pleurocythere, and others). Species of both groups demonstrate high evolutionary rates, and the duration of existence of many genera is restricted by the Jurassic period. The first group comprises only Middle Jurassic (Bajocian and Bathonian) taxa. About 30% of the taxa included in the second group appeared during the Early Jurassic, and the others (70%) originated during the Middle Jurassic. Nearly all of the taxa became extinct during the Late Jurassic; only 15% of them remained in existence in the Early Cretaceous, and one taxon (*Cytherelloidea*) has survived until the present. The lifespans of species in both groups generally corresponded to 5 - 10 m.y.

The third group comprises cosmopolitan taxa with a very wide geographic distribution (*Bairdia, Cytherella, Bythocypris, Paracypris, Schuleridea, Klieana, Monoceratina*, and *Bythoceratina*) that are long-lasting, i.e., they have a decelerated evolutionary rate. Many of them (with the exception of *Bairdia* and *Monoceratina* of Early Paleozoic origin) appeared during the Earliest Jurassic and have existed until the present; the two exceptions are *Klieana*, whose representatives became extinct at the end of the Cretaceous, and *Schuleridea*, which existed until the Neogene. The decelerated evolution of these taxa makes them useless for solving stratigraphic problems. Nevertheless, they are good ecology markers and provide important information on sedimentation environments.

The last group is indicative of specific habitat environments. These taxa of wide latitudinal range populated water basins with a considerable diapason of desalination. Their endemism, caused by the physicochemical parameters of the water basin, is notable at the family (Limnocytheridae and Darwinulidae) or superfamily (Darwinulacea) levels. In the Jurassic rocks of Syria, the family Limnocytheridae is represented by genera *Limnocythere, Theriosynoecum*, and *Bisulcocypris*; the family Darwinulidae is represented by the single genus *Darwinula*. These genera have a very great diversity of species. This complicates their usage in stratigraphy, because they can occur in lagoons, overdrying reservoirs, pools, lakes, and river deltas, i.e., in all basins where speciation is greatly dependent on the physical and chemical properties of the water. Genera *Limnocythere* and *Theriosynoecum* have existed since the Jurassic, *Bisulcocypris* has existed

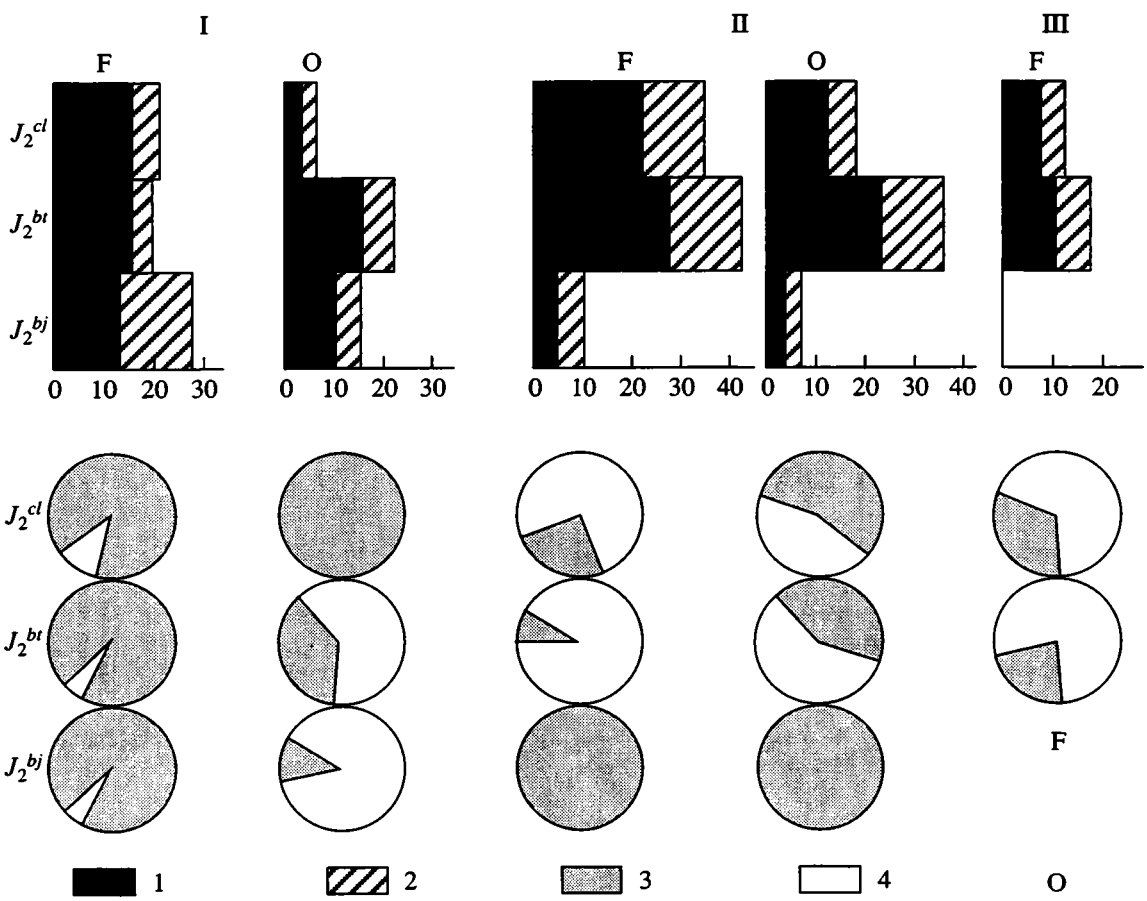


Fig. 3. Abundance of generic and specific taxa and proportions of their endemic and cosmopolitan forms in foraminifer and ostracod assemblages from the Middle Jurassic deposits of Syria.

I – The Anti-Lebanon; II – The Coastal Mountains; III – The Kurd-Dag, F – foraminifers; O – ostracods.

1 – genera; 2 – species; 3 – cosmopolites; 4 – endemics.

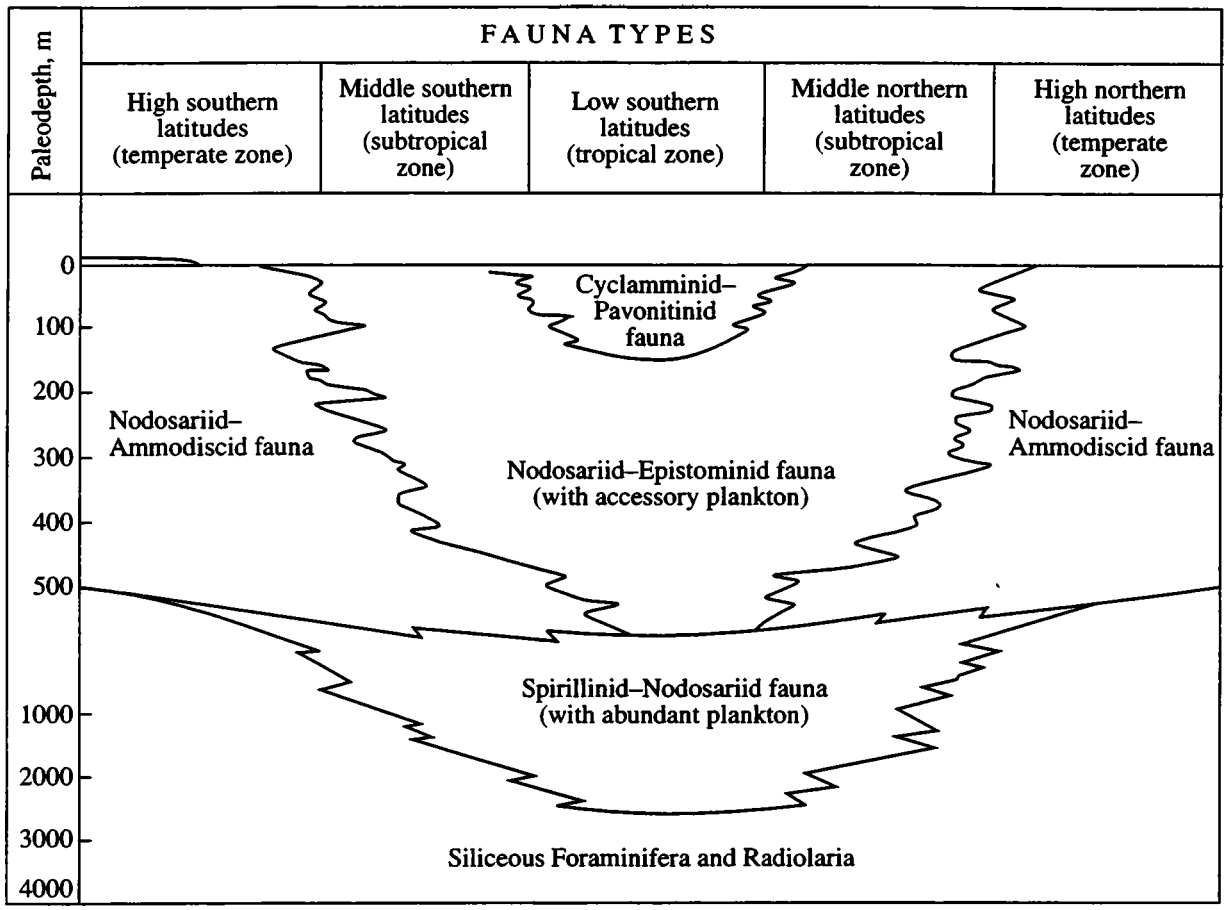


Fig. 4. Schematic latitudinal and bathymetric zonation in the distribution of Jurassic benthic foraminifers, after Basov (1991).

since the Triassic, and *Darwinula* forms are Devonian in origin. From the viewpoint of paleoecology, this group is very important, because all the above genera live in various modern basins.

Thus, it is evident that representatives of the first two groups are important for solving stratigraphic problems, whereas the third and fourth groups of fauna are useful for ecological analysis.

Below, we consider variations in the taxonomy and abundance of foraminifers and ostracods distinguished for the Middle Jurassic of Syria, because this stratigraphic interval bears the richest fauna, and corresponding sections in the Anti-Lebanon and Coastal Mountains are the most complete from a stratigraphic viewpoint. The Middle Jurassic of the Palmyrides and Kurd-Dag will be analyzed using only foraminifers, since ostracods have not been studied there.

THE ANTI-LEBANON

Aalenian and Bajocian Stages

The Aalenian–Bajocian sediments of southwestern Syria (the Anti-Lebanon), deposited in sublittoral basins and represented by organoclastic oolitic and clayey limestones with clay and marl interbeds, were the habitat for representatives of genera *Praelamarckina, Lamarckella, Reincholdella* (family Ceratobuliminidae), and *Epistomina* (family Epistiminidae); for abundant species of genera *Lenticulina* and *Citharina* (families Lenticulinidae and Vaginulinidae); and for lituolids, ataxophragmiids, and polymorphinids. All these taxa consist of widespread forms of the cosmopolitan boreal type. The Tethyan endemics are represented by scarce *Paleopfenderina gracilis* (Redmond) and *Timidonella sarda* Boussolet, Chabrier et Fourcade (Fig. 3, I), which are randomly distributed in the rock as single shells and as occasional mass accumulations. Amongst the 29 species of the 11 genera distinguished within this interval, 98% are known from eastern Europe (Dnieper–Donets depression, southwestern Donbass), the Crimea, Poland, and Egypt, and 2% have been described in Saudi Arabia, Morocco, and southwestern France. It is remarkable that all species available here are typical of marine basins with normal water salinity. The lowermost part of the Middle Jurassic sequence of the Anti-Lebanon is Aalenian–Bajocian in age, according to a recovered foraminifer assemblage with *Praelamarckina humilis* Kapt. It also

bears an ostracod assemblage with abundant shells of *Darwinula* sp., a form that evolves only under conditions of very low salinity and shallow depth (low-salinity lagoons, lakes, and river deltas). The middle part of the sequence consists of dolomites and is barren of ostracods. This part underlies a member of alternating limestones and clays. The poor ostracod assemblage from the basal strata of the member includes singular specimens of the cosmopolitan genus *Cytherella*. The genus belongs to the stenohaline type, and its modern representatives exist under a water salinity of 32 - 37‰ in shallow zones and 34 - 35‰ in deeper basin parts. Upward the section, this fauna is replaced by genera that occur only inside the Tethyan regions. Many of them are also known from Bajocian deposits of the Middle East (eastern Egypt, Jordan, and Israel; Basha, 1980, 1992; Rosenfeld *et al.*, 1987) enclosing genera of families Progonocytheridae (*Glyptocythere, Ektyphocythere*, and *Zergacythere*), Cytherideidae (*Schuleridea*), and Cytherelloidae (*Cytherelloidea*). This group is able to survive in slightly lower salinity. According to Neale (1988), the salinity range is 5 - 35‰ for *Schuleridea* and 10 - 35‰ for *Cytherelloidea* forms. This indicates that the top part of the sequence is dominated by endemic ostracod forms that are resistant to freshening, while cosmopolitan elements become less important here.

Bathonian Stage

In southwestern Syria, the Bathonian stage begins with an expanding marine transgression and basin deepening. However, local uplifts of that time retained adjacent shallow zones with local hiatuses in sedimentation. A lower part of the Bathonian sedimentary sequence near the village of Arneh (the Hermon Mountains in the Jebel Sheikh piedmonts) consists of organoclastic and oolitic limestones locally containing abundant shell detritus and accumulations of unsorted fauna. This part shows distinct lithologic features indicative of reworking and erosion unconformities. Upward, the deposits grade into oolitic, clayey, and pelitomorphic carbonates, which characterize sedimentation under deeper and quieter conditions in a setting remote from the coastal line. Its greater depth, free connections with the open sea, and normal water salinity are evidenced by the planktonic foraminifers *Globuligerina bathoniana* (Pazdro) and *G. meganomica* K. Kuznetsova that occur at this higher level. Shells of both species are accessory components of a peculiar and rich assemblage of benthic foraminifers represented mainly by the cosmopolitan genera *Ammobaculites, Ichthyolaria, Lenticulina, Planularia, Citharina, Eoguttulina, Lamarckella, Reincholdella*, and *Epistomina*. Associated Tethyan endemics consist of scarce shells of *Riyadhella hemeri* Redm. and *R. regularis* Redm.

In addition to foraminifers, the Arneh section yielded abundant ostracods from its lower and upper parts, while its middle part is free of this fauna. The lower ostracod assemblage includes genera *Ektyphocythere, Glyptogatocythere*, and *Virgulacytheridea*, which also are known from Middle Eastern sections (Oertli, 1966; Basha, 1980; Rosenfeld *et al.*, 1987). In general, this assemblage includes species occurring only in the southern boundary basins of the Tethys. The upper assemblage comprises rich and diverse ostracod fauna similar to that reported from the upper part of the Sheriff Formation (Jebel Maghara region in eastern Egypt; Rosenfeld *et al.*, 1987). The main components of the assemblage are genera of the family Progonocytheridae (*Procytheridea, Progonocythere, Fastigatocythere, Glyptocythere, Glyptogatocythere, Zergacythere, Ektyphocythere*, and others). Their species have wider distribution areas, e.g., 30% of them are known not only in the Middle East (Israel, Egypt, and Jordan), but also in western Europe (France, Germany, and Switzerland) and North America (Wyoming) (Peterson, 1954; Oertli, 1957, 1959, 1966; Hirsch, 1990). Abundant species of the *Cytherella* genus are dispersed throughout the entire upper sequence.

Callovian Stage

During the Callovian, the marine basin of southwestern Syria was similar to that of the late Bathonian. As before, it was mostly shallow with dominant carbonate sedimentation. The Callovian sediments in the southern Anti-Lebanon are as thick as 700 m and are represented by monotonous pelitomorphic limestones that experienced local dolomitization. The sequence also includes marl and clay interbeds, gradually replaced by limestone facies in northern areas. In addition to solitary and colonial corals, brachiopods, pelecypods, and gastropods, these sublittoral deposits contain both benthic and planktonic foraminifers. As in the preceding Bathonian assemblage, the Callovian foraminifers of the Anti-Lebanon are represented mainly by cosmopolitan boreal taxa (genera *Lituotuba, Verneuilina, Verneuilinoides, Paleogaudryina, Ophtalmidium, Lenticulina*, and *Planularia*), of low diversity and abundance. Only 18 species have been encountered among specimens of 13 genera distinguished in the Anti-Lebanon. Planktonic forms (*Globuligerina calloviensis* K. Kuznetsova) occur here as solitary specimens dispersed in carbonate deposits in the northern areas. They were identified in thin-sections of pelitomorphic limestones from the Wadi Al Karn and Wadi Fawuar sequences, which characterize the sublittoral (external infralittoral) zone of the basin. Associated Tethyan and sub-Tethyan forms also have low diversity and abundance. The five genera distinguished among them (*Nautiloculina, Praekurnubia, Kurnubia, Paleopfenderina*, and *Pseudocyclammina*) are represented by only six species (Plate I).

The ostracod assemblage from the Callovian deposits of the Anti-Lebanon is impoverished. Poorly preserved representatives of the *Cytherella* genus were recovered from a member of alternating clays and limestones in

the lower part of the Wadi Al Karn section. Among them we distinguished only *Cytherella index*, a species that is widespread in the Bathonian–Oxfordian deposits of the Middle East and has also been reported from the Oxfordian shelf sediments of Atlantic Canada (Rosenfeld *et al.*, 1987, 1991; Ascoli, 1976).

THE COASTAL MOUNTAINS

Considerably different distribution patterns of foraminifer and ostracod assemblages are typical of the studied Middle Jurassic sections of the Coastal Mountains. The first remarkable feature is the much higher abundance of Tethyan and sub-Tethyan forms, which dominated the foraminifer assemblages since the Bathonian time. The boreal-cosmopolitan foraminifers defined the assemblage type only during the initial Middle Jurassic, i.e., in the Bajocian, and then became subordinate in abundance and taxonomic diversity.

Bajocian Stage

In the southern Coastal Mountains, the Bajocian limestones contain *Lenticulina centralis* (Terquem), *L. veta* (Hoffman), and *L. protracta* (Born.). Their shells are not very abundant, probably because rock dolomitization resulted in the poor preservation of paleontological remains. Regardless, it is evident that sediments bearing these benthic forms were deposited within deeper basin zones that were more remote from the coast line than the sedimentation settings of the Anti-Lebanon rocks. Such zones correspond to epibathyal (circalittoral) environments.

In the lower part of the Jdaida section, the Bajocian deposits of the Coastal Mountains enclose abundant ostracod remains occurring above dolomitic limestones. These poorly preserved remains belong to the genus *Bairdia*, which is cosmopolitan. In modern basins, this stenohaline genus is unable to survive in a water salinity below 31 - 32‰ (Masumov and Bykovskaya, 1978).

Bathonian Stage

Since the Bathonian, the considered basin was thickly inhabited by benthic foraminifers consisting mostly of endemic Tethyan genera and species. One could say that this time was the peak of foraminiferal fauna endemism in the Mediterranean. Among the 47 species distinguished in the Bathonian deposits of the Coastal Mountains, 32 are Tethyan endemics. The assemblages here are dominated by the families Cyclamminidae (genera *Dhrumella, Everticyclammina,* and *Flabellaminopsis*), Haurandiidae (*Haurania, Amijiella*, and *Limognella*), Chrysalidinidae (*Riyadhella, Redmondoides, Riyadhoides*, and *Pseudomarssonella*), Paleopfenderinidae (*Paleopfenderina, Pfenderella, Sanderella, Satornia, Praekurnubia*, and *Kurnubia*), and Dictyoconidae (*Kilianina, Meyendorffina*, and *Paracoskinolina*). The shells are usually of high and sometimes of mass abundance. Many genera (*Riyadhella, Redmondoides,* and *Pseudomarssonella*) are diverse and include 5 - 7 species. Planktonic foraminifers occur sporadically. As previously, they are low in abundance and are represented by two species of the *Globuligerina* genus (*G. bathoniana* Pazdro and *G. meganomica* K. Kuzn.) associated with few or solitary epistominids and lenticulinids (Plate II).

Comparing the Bathonian foraminiferal assemblages from the Coastal Mountains with those from the coeval deposits of southwestern France (Pelissie *et al.*, 1984), one can note the similarity in their generic and specific compositions. French investigators related the habitat of these assemblages to internal and middle infralittoral (sublittoral) zones with the shallow and warm conditions favorable for the evolution and prosperity of larger foraminifers. It is remarkable that cooccurring fauna include various corals, stromatoporoids, bryozoan, pelecypods, and brachiopods. Thus, everything is indicative of a shallow and warm setting. From the complete absence of terrigeneous rocks and the very low content of siliciclastic admixture in sediments, it is evident that this assemblage was remote from the coast line. In sections of the Coastal Mountains, the stratigraphic interval under discussion is represented mainly by limestones, including varieties of shallow origin. Interlayers of clay and marl are usually thin and scarce; members with such rocks and limestone alternation are somewhat significant in a few localities (Jdaida and Bab Janneh).

The Bathonian ostracods from the Coastal Mountains are more diverse than their assemblages from the Anti-Lebanon. Genera and species are twice as high in number here and compose ostracod assemblages dominated, similar to foraminifers, by cosmopolitan forms in the western sections, where genera *Bairdia, Cytherella, Macrocypris*, and *Bythocypris* locally yielded mass accumulations. In the eastern localities, the ostracod assemblages are rich in forms typical of the Middle East.

Ostracods from the Braj, Bab Janneh, and Misyaf sections are represented mostly by species known from Israel, Jordan, and eastern Egypt. They are also similar to the assemblage distinguished in the upper Bathonian deposits of the Arneh locality of the Anti-Lebanon. In addition to forms characteristic of these localities, the Qadmous and Mareen the East sections yielded several species known from Europe: *Pleurocythere favosa* Triebel, *Pl. impar* Triebel, and *Fabanella sarda* Malz. The first two of these species occur in Germany, Poland, and Ukraine; the last form is recognized in the Bajocian–Bathonian deposits of northwestern Sardinia (Malz *et al.*, 1985; Triebel, 1951; Pyatkova and Permyakova, 1978). In addition to genera characteristic of normal marine conditions, the sections of Bab Janneh, Misyaf, Bchili, Mareen the East, Wadi Jannam, and

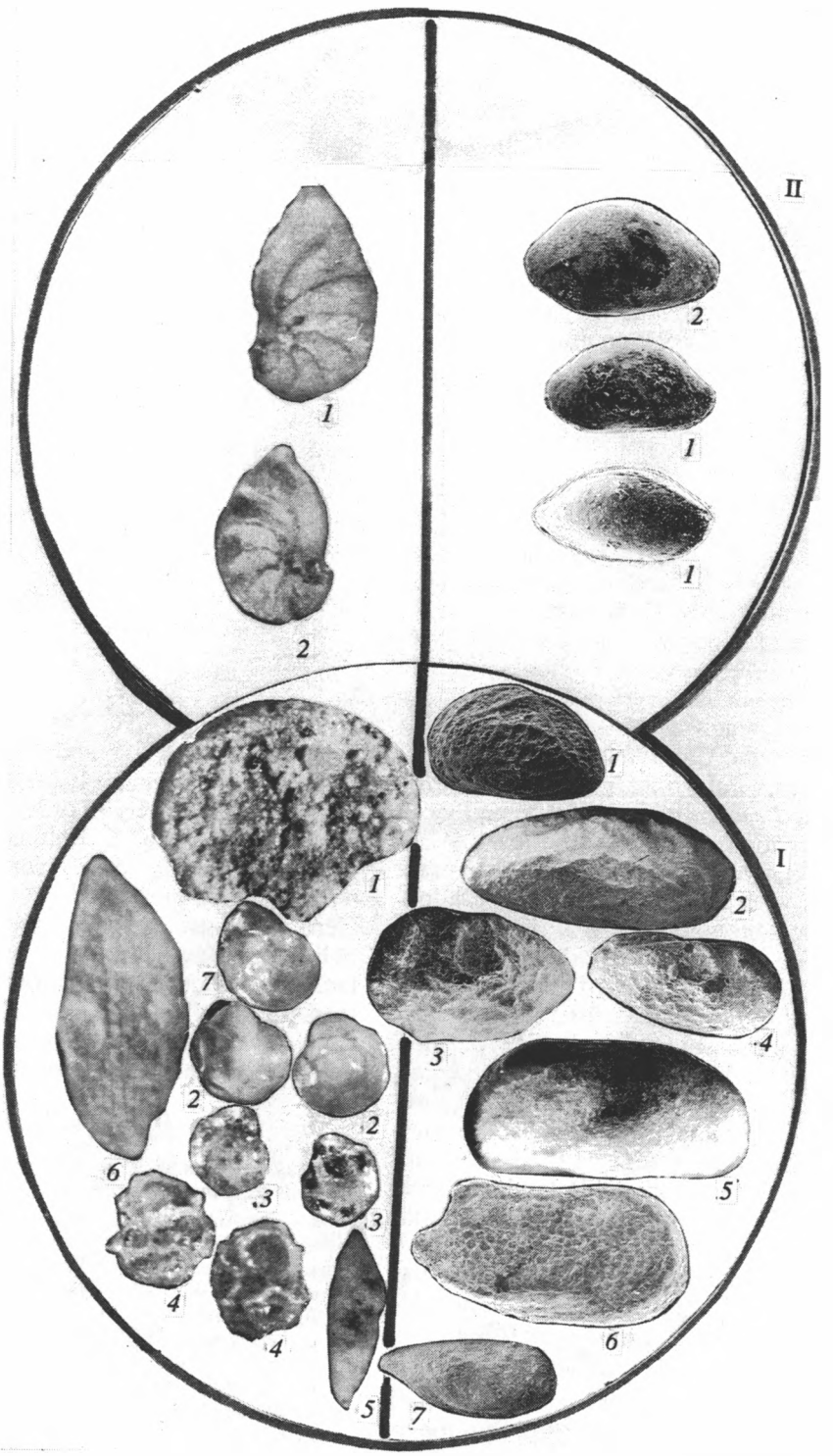


Plate I. Characteristic species of foraminifers and ostracods from the Bajocian deposits of Syria.

I – The Anti-Lebanon. Foraminifers: *1 – Timidonella sarda* Bouss., Chab. et Fourc., ×38; *2 – Epistomina* sp. 1, ×38; *3 – Lamarckella media* Kapt., ×75; *4 – Epistomina coronata* Terq., ×75; *5 – Citharina ksibiensis* Sossip. et Rohh., ×38; *6 – C. naourica* Sossip. et Rohh., ×38; *7 – Lenticulina varians* (Born.), ×38. Ostracods: *1 – Ektyphocythere shulamitae* Rosenf., Gerry et Honigst., ×56; *2 – Ekt. dierallaensis* Basha, ×56; *3 – Zergacythere* aff. *subiehiensis* Basha, ×60; *4 – Glyptocythere huniensis* Basha, ×53; *5 – Darwinula* sp., ×53; *6 – Cytherelloidea* sp., ×53; *7 – Monoceratina* sp., ×56.

II – The Coastal Mountains. Foraminifers: *1, 2 – Lenticulina polymorpha* (Terq.), ×75. Ostracods: *1 – Bairdia* sp. 1, ×26; *2 – Bairdia* sp. 2, ×26.

Qadmous bear endemic freshwater forms (*Theriosynoecum* and *Darwinula*) occurring in basal horizons.

The Bathonian ostracod assemblage from the Wadi Jannam locality is remarkably different from coeval fauna characterizing other sections of the Coastal Mountains, the Anti-Lebanon, and the Middle East. Only cosmopolitan species are common for all of them. Many forms (*Macrodentina tenuistriata* Malz, M. *whatleyi* Kilenyi, and others) occurring near the Wadi Jannam are known from western Europe (Malz, 1958; Kilenyi, 1978). The upper part of this section yielded only mass accumulations of cosmopolitan taxa (*Bairdia* and *Cytherella* genera).

Callovian Stage

During the Callovian, the marine basin of western Syria retained its former features. It was a shallow warm sea located on a carbonate platform, favorable for the evolution of larger foraminifers representing endemic Tethyan fauna. Because of stable life conditions, this fauna inherited many of the peculiarities of its predecessors; however, several Bathonian genera (*Meyendorffina, Kilianina, Haurania, Protopeneroplis,* and *Paracoskinolina*) reached the terminal stage of their evolution during the Callovian. Other genera of that time (*Kurnubia, Paleopfenderina, Riyadhella, Redmondoides*, and *Pseudomarssonella*) experienced significant transformations at the specific level. Western areas of the basin, which opened toward the present Mediterranean Sea, were habitats for different Callovian assemblages that consisted mostly of cosmopolitan forms (Plate III). For example, the Callovian deposits of the Bchili and Braj sections revealed peculiar assemblages of *Paleogaudryina magharaensis* Said et Barakat, *Epistomina depressa* Said et Barakat, *E. parastelligera* (Hofner), *Eoguttulina polygona* Terq., *E. althi* Lloyd, *E. metensis* (Terq.), *E. anglica* Cushm. et Ozawa, and others. Both the generic and specific compositions of such assemblages suggest their similarity to coeval assemblages from England, Poland, and Germany. In addition, they contain species reported from the Callovian deposits of the Sinai Peninsula, where dominant forms are represented, as in the Bajocian and Bathonian faunas, by widespread cosmopolitan elements, and nearly all Tethyan endemics are missing.

Ostracods characterizing the Callovian sediments of the Coastal Mountains were encountered in the Wadi Chkeir, Mareen the East, Bab Janneh, and Bchili localities. The assemblages consist of endemics in the first two localities, and they are rich in cosmopolitan forms in the second two.

Thin clay interbeds in the middle part of the Wadi Chkeir section contain mass shell accumulations of the freshwater genera *Theriosynoecum* and *Bisulcocypris*. In the section of Mareen the East, similar clay interbeds above the boundary with the Bathonian deposits bear abundant ostracods of the genus *Klieana*, characteristic of low-salinity conditions. Presumably, a short period of river runoff affected the environment in the Callovian marine basin of the central Coastal Mountains and provided for the origin of the corresponding ostracod assemblages. In other areas, similar events are undetectable.

The Callovian ostracod assemblages from the Bchili and Bab Janneh localities show features characteristic of both Middle Eastern and western European regions. They are rich in either widespread or cosmopolitan species. In the Bchili section, this Callovian fauna is similar to coeval ostracod assemblages from the Hamakhtesh Hagadol and Coastal localities of Israel (Rosenfeld and Honigstein, 1991). In addition to European species, it also contains forms known in Tanzania, Madagascar, Saudi Arabia, Jordan, Lebanon, Israel, and Egypt.

Thus, the Callovian basins of western Syria were habitats of various foraminifer and ostracod assemblages: thermophilic Tethyan endemics populated the main areas, while cosmopolitan forms were localized in basins of the western Coastal Mountains.

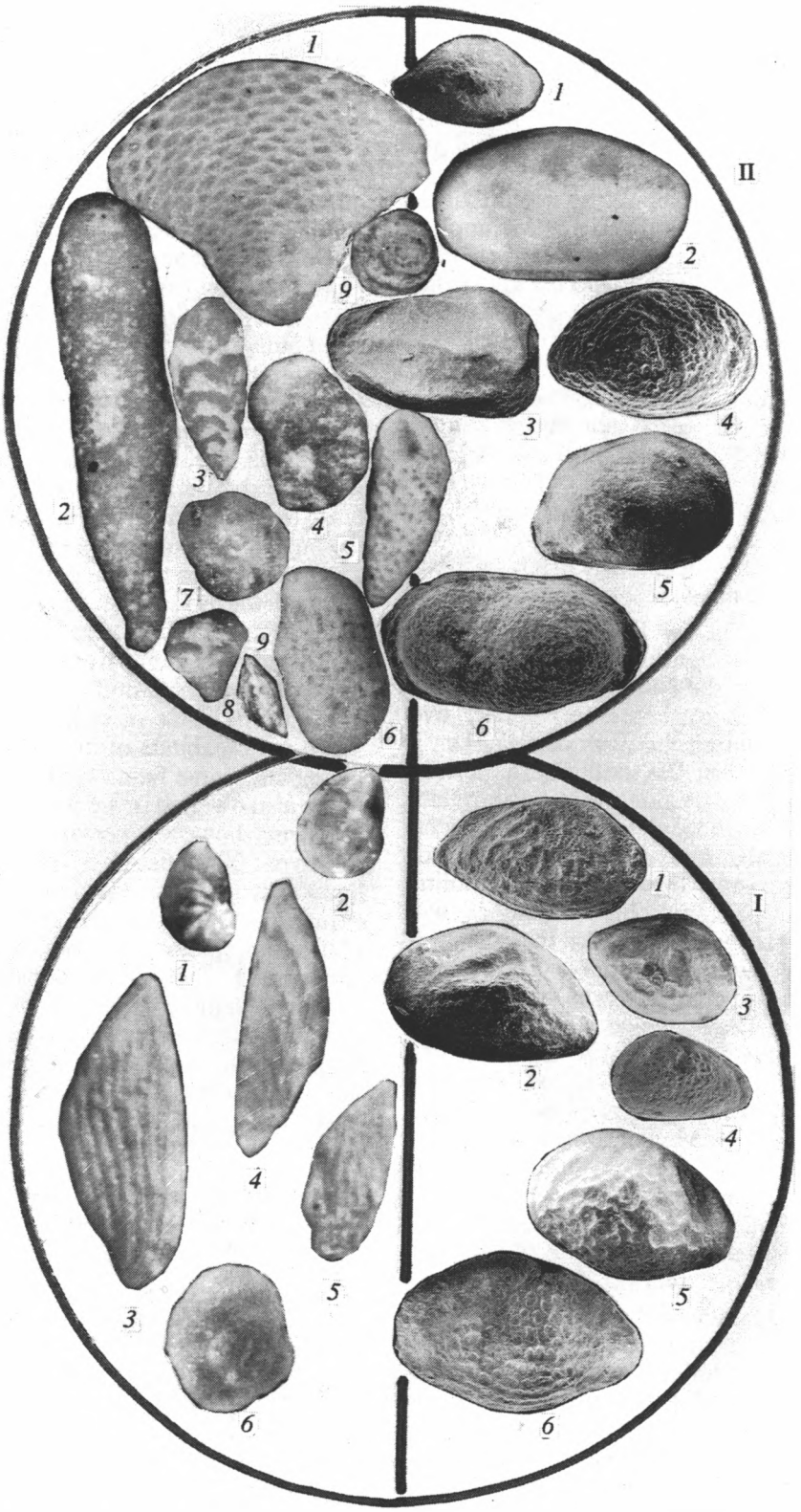
THE PALMYRIDES AND THE KURD-DAG

In the Palmyrides and the Kurd-Dag, the Middle Jurassic deposits have a limited distribution, and we will consider them briefly in order to complete our analysis of the foraminiferal assemblages of Syria. In the Kurd-Dag, northwestern Syria, foraminifers were encountered in Bathonian and some Callovian sediments. Here, this stratigraphic interval is represented by oolitic, pseudooolitic, and pelitomorphic limestones lacking carbonate-terrigenous interbeds. The limestones yielded a foraminiferal assemblage with dominant Tethyan endemics typical of the Bathonian Stage: genera *Dhrumella, Protopeneroplis, Haurania*, and

Plate II. Characteristic species of foraminifers and ostracods from the Bathonian deposits of Syria.

I – The Anti-Lebanon. Foraminifers: *1 – Lenticulina polymorpha* (Terq.), ×38; *2 – L. varians* (Born.), ×38; *3 – Citharina chlatrata* (Terq.), ×38; *4 – C. pseudoflabellata* Sossip. et Rohh., ×38; *5 – C. heteropleura* (Terq.), ×38; *6 – Globuligerina bathoniana* (Pazdro), ×38. Ostracods: *1 – Ektyphocythere triangula* (Brand), ×56; *2 – Glyptogatocythere malzi* Basha, ×53; *3 – Glyptocythere huniensis* Basha, ×53; *4 – Fastigatocythere bakeri* Basha, ×53; *5 – Progonocythere honigsteini* Rosenf. & Gerry, ×53; *6 – Procytheridea crassa* Peterson, ×53.

II – The Coastal Mountains. Foraminifers: *1 – Kilianina blancheti* Pfender, ×38; *2 – Haurania deserta* Hens., ×38; *3 – Riyadhella regularis* Redm., ×38; *4 – Dhrumella evoluta* Redm., ×38; *5 – Praekurnubia crusei* Redm., ×38; *6 – Meyendorffina bathonica* Aurouze et Bizon, ×38; *7 – Epistomina turgidula* Pazdro, ×38; *8 – Verneuilinoides jannamensis* Grig. et K. Kuzn., ×38; *9 – Trocholina conosimilis* Subb. et Srivast., ×38. Ostracods: *1 – Bairdia* aff. *hilda* Jones, ×26; *2 – Cytherella index* Oertli, ×53; *3 – Pleurocythere impar* Triebel, ×53; *4 – Ektyphocythere shulamitae* Rosenf. & Gerry, ×60; *5 – Praeschuleridea hornei* Rosenf. & Gerry, ×53.


II
I

Paleopfenderina. In addition, it includes accessory specimens of Hauerinidae (*Pseudosigmoilina* and *Sigmoilina*). The assemblage was identified in thin-sections and can hardly be used in a quantitative comparison with other assemblages extracted from soft clayey rocks. A similar assemblage, though impoverished in composition, was distinguished in the Callovian clayey and pelitomorphic limestones of the Rajo and Smalek sections in the Kurd-Dag. It should be noted that the Bathonian sequence of the Kurd-Dag is incomplete, being transgressively overlapped by Lower Cretaceous deposits. Callovian units with a limited thickness (less than 3 - 4 m) are distinguishable only in some sections, where they unconformably rest on Bathonian rocks and are also overlain by Lower Cretaceous deposits above a sharp erosion boundary. Lithologic features of the rocks, which lack a terrigeneous admixture, suggest their deposition in a shallow basin; this is also evident from identified Tethyan endemic fauna of benthic foraminifers of the Bathonian and Callovian stages. French scientists (Pelissie *et al.*, 1984) believed that similar assemblages from southwestern France inhabited the internal and middle infralittoral (sublittoral) in that region.

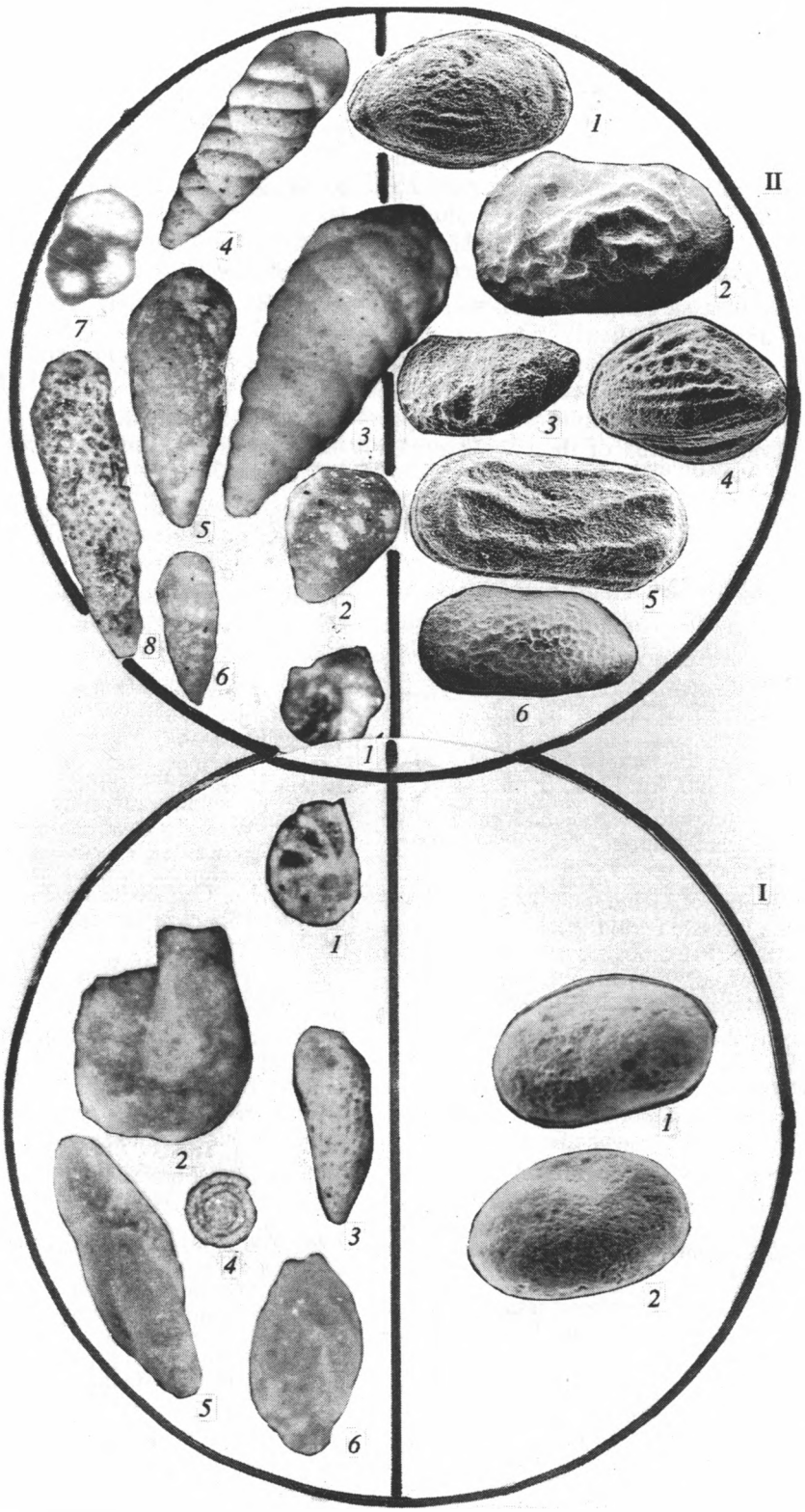
The mountain range of the Palmyrides in central Syria (Fig. 2) is of special significance when analyzing the geological history of the Jurassic marine basin in this region. The marine transgressions spread over these mountains only during the peak periods, i.e., in the Bathonian and Oxfordian. We studied corresponding marine deposits in several sections of the Palmyrides, among which the Al Zbeidi locality, comprising Bathonian sediments, was the most interesting. The sediments here are represented by organoclastic, oolitic, stromatoporoid, and pelitomorphic limestones with numerous beds of terrigeneous rocks that occur as members alternating with limestones. The terrigeneous material, abundant erosion marks, and lithologic features of the limestones in the section suggest their deposition in a basin zone adjacent to the coast line. The rich and diverse foraminifer assemblage from these rocks includes the following abundant Tethyan endemics: *Nautiloculina oolithica* Mohler, *Dhrumella evoluta* Redm., *Haurania deserta* Henson, *Amijiella amiji* (Henson), *Kilianina blancheti* Pfender, *Redmondoides lugeoni* (Sept.), *R. primitivus* (Redm.), and others. Only *Paulina paula* (Pazdro) and some species of the genus *Ammobaculites* are of a different origin.

As follows from the data synthesis above, foraminifer assemblages from different zones of the Jurassic paleobasins in the Eastern Mediterranean significantly varied in their composition, taxonomic diversity, and abundance (Plates I - III). An inferred succession of coastal to deep-water sedimentary facies bearing foraminifers of a single tropical climatic zone was published by Basov and Vasilenko (1986). These researchers concluded that larger foraminifers with a complex internal structure (Pavonitinidae, Orbitolinidae, Cyclamminidae, and Alveolinidae) associate with shallow facies of the neritic zone. In addition, miliolids, larger nodosariids, epistominids, and trocholinids are also characteristic of this zone. A deeper part of the neritic zone (lower sublittoral and external shelf) is a habitat for nodosariids, epistominids, ceratobuliminids, and ataxophragmiids. Basov (1991) suggested the most illustrative scheme of distribution for benthic foraminifers according to latitudinal and bathymetric zonation (Fig. 4). However, this scheme is on a global scale, and when we observe a rapid lateral replacement of one fauna type by another within the coeval deposits of a single latitudinal and climatic zone, in our case, tropical, the paleoclimatic zonation can hardly be considered as the responsible factor. The observed replacement of assemblages and faunal groupings, which generally corresponds to the latitudinal-bathymetric succession of foraminiferal fauna types after Basov, seems to have been caused by a similar phenomena of lesser significance. The depth and temperature variations in our case were considerably less significant, such as they should be in a single basin or even in a part of the basin. We suggest the following distribution model for the studied Middle Jurassic fauna. Within the low latitudes (tropical zone), shallow and warm sea areas above the carbonate platform (0 - 100 m, i.e., the littoral and upper sublittoral) were populated by endemic tropical assemblages of foraminifers (the cyclamminid-pavonitinid fauna type, after Basov). The areas of temperate latitudes (subtropical zone) were habitats of nodosariid-epistominid assemblages (the same fauna type, after Basov). This Syrian fauna also dwelled in the warm sea above the carbonate platform; however, seaward, in remote and deeper (down to 500 m) basins of the lower sublittoral and epibathyal zones, this fauna lived in somewhat lower temperatures. As follows from the distribution analysis, the nodosariid-epistominid assemblages were the most widespread. They can be considered as cosmopolitan, and they demonstrate a bipolar trend in foraminifer evolution during the Jurassic period. The Jurassic basins of the eastern Mediterranean were free of both nodosariid-ammodiscid fauna, which is characteristic of the temperate zone at higher latitudes, and the spir-

Plate III. Characteristic species of foraminifers and ostracods from the Callovian deposits of Syria.

I – The Anti-Lebanon. Foraminifers: *1 – Lenticulina polonica* (Wisn.), ×38; *2 – Lituotuba nodus* Kosyr., ×38; *3 – Praekurnubia crusei* Redm., ×38; *4 – Spirillina eichbergensis* Kueb. et Zwingli, ×38; *5 – Eoguttulina triloba* (Terq.), ×75; *6 – E. oolithica* (Terq.), ×75. Ostracods: *1 – Cytherella index* Oertli, ×56.

II – The Coastal Mountains. Foraminifers: *1 – Conorboides scutuliformis* Seib., ×38; *2 – Redmondoides lugeoni* (Sept.), ×38; *3 – Paleopfenderina salernitana* (Sart. et Cresc.), ×38; *4 – Kurnubia wellingsi* Henson, ×38; *5 – K. bramkampi* Redm., ×38; *6 – Riyadhella intermedia* Redm., ×38; *7 – Globuligerina calloviensis* K. Kuzn., ×56; *8 – Kurnubia variabilis* Redm., ×38. Ostracods: *1 – Progonocythere* aff. *paractilla* Whatley, ×60; *2 – Afrocythere faveolata* Bate, ×49; *3 – Bythoceratina* sp., ×64; *4 – Ektyphocythere zoharensis* Rosenf. & Gerry, ×60; *5 – Cytherelloidea aazourensis* Bishoff, ×64; 6 – *Mandelstamia* sp., ×53.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ilinid-nodosariid assemblages, which populated the bathyal areas (depth range 500 - 3000 m) between the equatorial and high-latitude zones.

The analyzed patterns of benthic foraminifer distribution in the Jurassic basins of Syria show that bathymetric and related thermal factors were the most important for their dispersal.

The salinity of the paleobasin in the eastern Mediterranean during the Jurassic followed the patterns depicted by a study of recent foraminifers. The studied Jurassic sections of Syria indicate that, in coastal basin areas, the water was freshening due to river runoff.

Ostracods from the Jurassic basins of Syria show the same bathymetric and areal distribution patterns. Endemic Tethyan ostracod assemblages inhabited the inner shelf and shallow littoral; the deeper outer shelf, which opened toward the sea (sublittoral and epibathyal), was the habitation area of their widespread cosmopolitan forms.

Reviewer V.A. Basov

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Radiolarians and Stratigraphy of the Campanian–Maestrichtian Deposits in Southwestern Cyprus

L. G. Bragina and N. Yu. Bragin

Institute of Geology, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia

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Abstract – Studied radiolarians characterize the Upper Cretaceous deposits of southwestern Cyprus, namely, the Kannaviou (bentonite clays, up to 100 m in thickness) and Kathikas (gravity mixtites, 100 - 200 m in thickness) formations. The radiolarian *Amphipyndax pseudoconulus* Zone of the Campanian is distinguished in the lower part of the Kannaviou Formation sequence. Its upper part and the Kathikas Formation contain a radiolarian assemblage of the *Amphipyndax tylotus* Zone of the upper Campanian–Maestrichtian. New data facilitate a more refined age determination of the deposits, which were previously assigned to the Maestrichtian.
Key words: *Upper Cretaceous, Kannavion and Kathikas formations, stratigraphy, radiolarians.*

The Upper Cretaceous deposits of Cyprus are of considerable interest to regional geology, because they form the basal neoautochthon strata and directly overlie the allochthonous formations of the Mamonia Complex and ophiolites of the Troodos massif. Information on such basal horizons provides insight into the history of the formation of the geological structure of Cyprus, yet their stratigraphy is poorly studied. The deposits are barren of macrofauna and contain scarce remains of foraminifers. Only radiolarians, which are thus far insufficiently studied, occur there in abundance.

The Upper Cretaceous deposits of southwestern Cyprus are subdivided into the following units: the Turonian–early Campanian Perapedhi Formation, consisting of metalliferous umbers and cherts with radiolarite interbeds resting on the eroded surface of the upper pillow-lavas of the Troodos massif (Blome and Irwin, 1985; Bragin and Bragina, 1991; Thurow, 1991); the Kannaviou Formation, consisting of bentonite clays with sandstone and acid tuff interlayers; the Kathikas Formation, consisting of olistostromes with a clayey matrix; and the Lefkara Formation with the Maestrichtian basal beds composed of limestones and marls. The Kannaviou and lower Kathikas deposits, which contain the richest radiolarian assemblages (Plates I - III), were the primary interest of this study. They were investigated in two sections, which are described below.

In the lower course of the Mavrokolimbos River, 1 km above its mouth, there are outcrops of pillow-lavas overlain by Upper Cretaceous deposits. From the base upward, this section includes the following beds (Fig. 1):

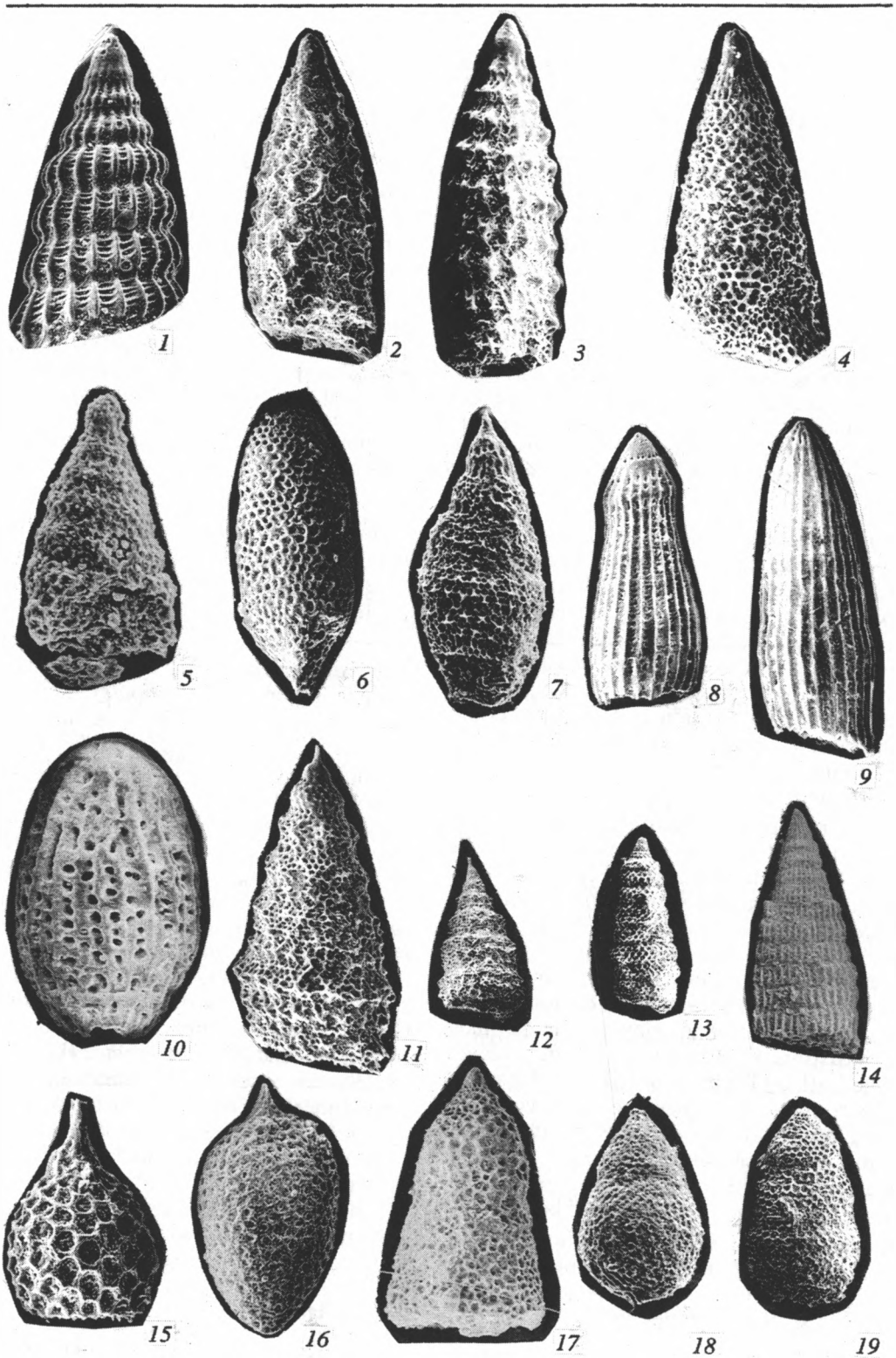
(1) Weathered pillow-basalts. Apparent thickness: 1 m.

(2) Clays that are reddish to brown or dark brown, laminated, sometimes platty, and bear lenses of yellow siliceous umbers. At the base, they fill pockets on the eroded surface of the basalts and include basalt pebbles. Thickness: 0 - 1 m.

(3) Bentonite clays that are greenish-grey, massive, poorly laminated, sometimes platty, and bear interlayers of grey platty shale (20 - 30 cm) and siltstone (up to 1 cm). In the basal and higher parts of the bed, the following radiolarians occur: *Afens lirioides* Reidel et Sanfilippo, *Alievium gallowayi* (White), *A.* aff. *riedeli* Pessagno, *A.* sp., *Amphipyndax pseudoconulus* (Pessagno), *A. stocki* (Campbell et Clark), *Archaeodictyomitra regina* (Campbell et Clark), *Crucella* sp., *Dictyomitra duodecimcostata* (Squinabol), *D. formosa* (Squinabol), *D. koslovae* Gorbovetz, *D. torquata* Foreman, *Halesium sexangulum* Pessagno, *Patellula enessceei* Empson-Morin, *P. verteroensis* Pessagno, *Patulibracchium* sp., *Praeconocaryomma californica* Pessagno, *P. lipmanae* Pessagno, *P. universa* Pessagno, *Pseudoaulophacus parguerensis* Pessagno, *P. venadoensis* Pessagno, *Phopalosyrigium* sp., *Stichomitra* ex gr. *alamedaensis* (Campbell et Clark), *Thanarla veneta* (Squinabol), *Theocampe urna* (Foreman), and *Xitus asymbatos* (Foreman). Specimens of *Alievium superbum* (Squinabol), *Crucella messinae* Pessagno, *Dictyomitra multicostata* Zittel, *Pseudoaulophacus praefloresensis* Pessagno, *Rhopalastrum* cf. *tumidum* Lipman, and *Stichomitra cechena* Foreman appear at a level 0.5 m above the base. Another level 1.2 m above the base also bears *Amphipyndax tylotus* Foreman and *Hemicryptocapsa* cf. *conara* Foreman, but species such as *Alievium superbum* (Squinabol), *Crucella messinae* Pessagno, and *Theocampe urna* (Foreman) disappear here. Thickness: 1.8 m.

(4) Clays that are greenish-grey and massive with unclear lamination, comprising frequent interbeds (20 - 30 cm) of grey platty opokalike claystones and scarce lensing interlayers of white flocculent tripoli. At a level 4 m above the base, the bed contains

Plate I



Phaseliforma laxa Pessagno and *Pseudoaulophacus floresensis* Pessagno. Higher (5.5 m above the base), it bears *Crucella irwini* Pessagno and *Rhopalastrum tumidum* Lipman, but *Phaseliforma laxa* Pessagno disappears here. *Archaeospongoprunum bipartitum* Pessagno, *Clathropyrgus titthium* Reidel et Sanfilippo, *Crucella* cf. *cachensis* Pessagno, *C. latum* (Lipman), *Prunobracchium longum* Pessagno, and *Rhopalastrum* ex gr. *ingens* Lipman occur 6.5 m above the base, whereas *Dictyomitra multicostata* Zittel disappears here. The first occurrences of *Acaeniotyle* ex gr. *diaphorogona* Foreman, *A.* ex gr. *umbilicata* (Rust), and *Archaeospongoprunum venadoensis* Pessagno were noted 8 m above the base. This level is also marked by the disappearance of many of the species that occurred below: *Archaeospongoprunum bipartitum* Pessagno, *Crucella* cf. *cachensis* Pessagno, *C. latum* (Lipman), *Prunobracchium longum* Pessagno, and *Rhopalastrum* ex gr. *ingens* Lipman. The next level, 10 m above the base, revealed occurrences of *Acidnomelos* sp. and *Phaseliforma carinata* Pessagno and the absence of *Acaeniotyle* ex gr. *diaphorogona* Foreman, *Rhopalastrum tumidum* Lipman, and several forms characteristic of Bed 3 [*Crucella* cf. *messinae* Pessagno, *Dictyomitra duodecimcostata* (Squinabol), *Halesium* ex gr. *sexangulum* Pessagno, and *Xitus asymbatos* (Foreman)]. The upper level at 10.5 m marks the first occurrences of *Histiastrum aster* Lipman, *Hemocryptocapsa conara* Foreman, and *Theocampe* ex gr. *ascalia* Foreman as well as the disappearance of *Acaeniotyle* ex gr. *umbilicata* (Rust), *Amphipyndax tylotus* Foreman, *Archaeospongoprunum venadoensis* Pessagno, *Clathropyrgus titthium* Reidel et Sanfilippo, *Phaseliforma carinata* Pessagno, *Pseudoaulophacus praefloresensis* Pessagno, and *Stichomitra* ex gr. *alamedaensis* (Campbell et Clark). Thickness: 13 m.

The upper part of the Kannaviou Formation, together with the overlying Kathikas Formation, is exposed near the village of Kannaviou, where the section has following units (Fig. 1):

(1) *The Kannaviou Formation.* Clays that are greenish-grey and massive, enclosing interlayers of light grey opokalike claystone and greenish-grey tuffaceous siltstones. A 2-m-thick horizon in the lower part of the outcrop consists of white and light grey massive lithoclastic tuffs. Rocks above this horizon yielded the following radiolarians: *Afens lirioides* Reidel et Sanfilippo, *Amphipyndax pseudoconulus* (Pessagno), *A. stocki* (Campbell et Clark), *A. tylotus* Foreman, *Alievium gallowayi* (White), *A. superbum* (Squinabol), *Artostrobium* sp., *Crucella irwini* Pessagno, *Dictyomitra andersoni* (Campbell et Clark), *D. densicostata* Pessagno. *D. duodecimcostata* (Squinabol), *D. koslovae* Gorbovetz, *Euchitonia santonica* Lipman, *Phaseliforma laxa* Pessagno, *Praeconocaryomma californiaensis* Pessagno, *P. universa* Pessagno, *Pseudoaulophacus parguerensis* Pessagno, *P. praefloresensis* Pessagno, *Phopalosyrigium* sp., *Stichomitra* cf. *campi* (Campbell et Clark), and *S.* ex gr. *cechena* Foreman. Apparent thickness: 30 m.

(2) *The Kathikas Formation.* Mixtites with a matrix consisting of brick-red sandy unlaminated clays. Inclusions are unrounded, unsorted, and variable in size: from dominant small (1 - 10 cm) to individual large (1 - 5 m) blocks. They are composed mainly of rocks of the Mamonia Complex: Vlambouros and Akamas sandstones, siliceous rocks of the Episkopi Formation, and volcanic rocks and limestones of the Dhiarizos Group. The lower boundary of the Kathikas Formation is sharp and irregular and has pockets. In the basal part of the formation, lenses (up to 5 m in thickness) of bentonite clays, similar in composition to those from the Kannaviou Formation, bear the following radiolarians: *Acaeniotyle* ex gr. *diaphorogona* Foreman, *Afens lirioides* Reidel et Sanfilippo, *Alievium* sp., *Amphipyndax pseudoconulus* (Pessagno), *A. stocki* (Campbell et Clark), *Clathropyrgus titthium* Reidel et Sanfilippo, *Crucella latum* (Lipman), *Dictyomitra* ex gr. *duodecimcostata* (Squinabol), *Hemicryptocapsa conara* Foreman, *Phaseliforma laxa* Pessagno, *Pseudoaulophacus floresensis* Pessagno, *P. parguerensis* Pessagno, *P. praefloresensis* Pessagno, *Rhopalastrum tumidum* Lipman, *Septinastrum dogeli* Gorbovetz, *Stylosphaera* ex gr. *pusilla* Campbell et Clark, and *Xitus asymbatos* (Foreman). Thickness: 100 - 150 m.

(3) *The Lefkara Formation.* Limestones and marls that are white or light grey, platty and cloddish, and bear foraminifers of the late Maestrichtian (according to determinations by V.A. Krasheninnikov). Apparent thickness: 5 - 10 m. Higher, the section consists of calcareous Cenozoic deposits.

Thus, the Kannaviou and basal Kathikas deposits contain representative radiolarian assemblages that provide

Plates I - III. Characteristic radiolarian species from the Mavrokolimbos section (Magnification ×150, unless otherwise indicated).

Plate I. *1* – *Dictyomitra formosa* Squinabol, Sample 36-8, *tylotus* zone; *2* – *Amphipyndax* aff. *pseudoconulus* (Pessagno), Sample 36-8, *tylotus* zone; *3* – *Amphipyndax pseudoconulus* (Pessagno), Sample 36-8, *tylotus* zone; *4* – *Amphipyndax tylotus* Foreman, Sample 36-8, *tylotus* zone; *5* – *Amphipyndax stocki* (Campbell et Clark), Sample 36-10, *tylotus* zone; *6* – *Stichomitra livermorensis* (Campbell et Clark), Sample 36-10, *tylotus* zone; *7* – Theoperidae gen. et sp. indet., Sample 36-8, *tylotus* zone; *8* – *Thanarla veneta* (Squinabol), Sample 36-2, *pseudoconulus* zone; *9* – *Archaeodictyomitra squinaboli* Pessagno, Sample 36-2, *pseudoconulus* zone; *10* – *Artostrobium* sp., Sample 36-2, ×250, *pseudoconulus* zone; *11* – *Xitus asymbatos* (Foreman), Sample 36-2, *pseudoconulus* zone; *12* – Same specimen as in *11*, ×75; *13* – *Stichomitra* sp., Sample 36-10, ×75, *tylotus* zone; *14* – *Dictyomitra multicostata* Zittel, Sample 36-10, *tylotus* zone; *15* – *Rhopalosyringium* sp., ×250, *pseudoconulus* zone; *16* – Theoperidae(?) gen. et sp. indet., Sample 36-8, *tylotus* zone; *17* – Theoperidae gen. et sp. indet., Sample 36-2, *pseudoconulus* zone; *18* – *Stichomitra* aff. *campi* (Campbell et Clark), Sample 36-8, ×100, *tylotus* zone; *19* – *Stichomitra campi* (Campbell et Clark), Sample 36-8, ×100, *tylotus* zone.

Plate II

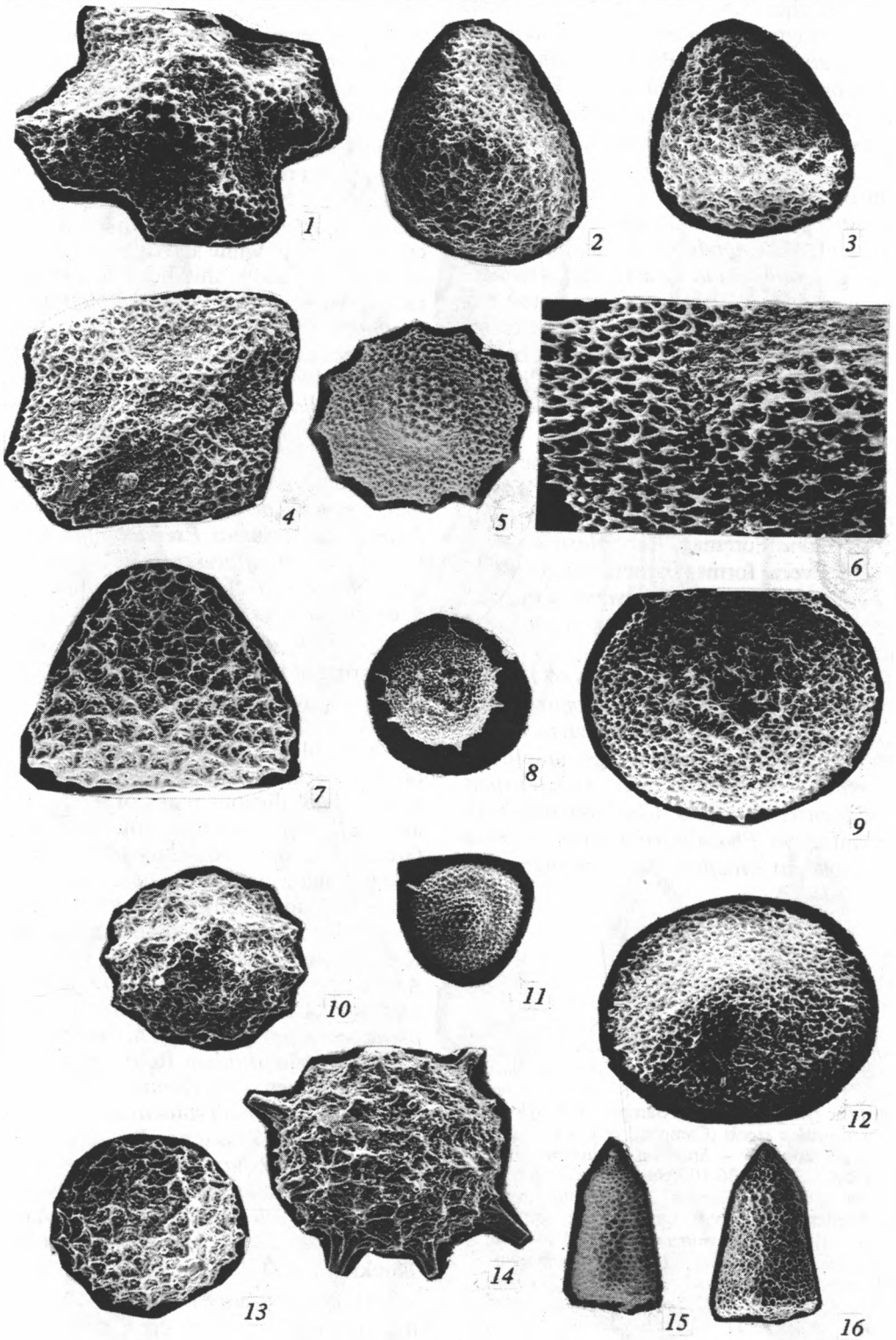


Plate II. *1, 4 – Crucella robusta* Bragina, sp. nov., Sample 36-8, *tylotus* zone; *2, 3 – Alievium*(?) sp., Sample 36-2, *pseudoconulus* zone; *5 – Pseudoaulophacus parqueraensis* (White), Sample 36-2, *pseudoconulus* zone; *6* – Same specimen as in *5*, ×400; *7 – Alievium gallowayi* (White), Sample 36-8, *tylotus* zone; *8 – Patellula enessceei* Empson-Morin, Sample 36-2, ×75, *pseudoconulus* zone; *9, 12 – Patellula verteroensis* (Pessagno), Sample 36-2, ×200, *pseudoconulus* zone; *10 – Praeconocaryomma universa* Pessagno, Sample 36-2, *pseudoconulus* zone; *11 – Pseudoaulophacus floresensis* Pessagno, Sample 36-8, ×75, *tylotus* zone; *13 – Praeconocaryomma californiaensis* Pessagno, Sample 36-8, *tylotus* zone; *14 – Praeconocaryomma* (?) sp., Sample 36-2, ×200, *pseudoconulus* zone; *15, 16 – Stichomitra* (?) sp., Sample 36-2, ×75, *pseudoconulus* zone.

insight into their stratigraphy. In the tropical oceanic regions, the Campanian–Maestrichtian interval is currently subdivided into the *Amphipyndax pseudoconulus* (Campanian) and *Amphipyndax tylotus* (late Campanian–Maestrichtian) zones of the radiolarian scale (Sanfilippo and Reidel, 1985). These zones are also distinguishable in the Cyprus sections.

We identified the *Amphipyndax pseudoconulus* Zone in the basal part of Bed 3 of the Mavrokolimbos section. Considering the occurrences here of *Afens lirioides, Alievium gallowayi, Dictyomitra koslovae*, and *Pseudoaulophacus parqueraensis*, one could suggest that, in this particular section, we are dealing with an upper part of the oceanic *pseudoconulus* zone, which is characterized by these species (Sanfilippo and Reidel, 1985). However, it also contains *Theocampe urna* and *Alievium superbum*, neither of which are typical of the upper *pseudoconulus* subzone (Sanfilippo and Reidel, 1985). In the Mavrokolimbos section, the zone has a very limited thickness (up to 1.2 m only) and may represent a condensed succession, which would explain the co-occurrence of species that are different in age. However, it is not inconceivable that the stratigraphic intervals of the above species in the section under consideration are somewhat different from those in oceanic successions.

The base of the *Amphipyndax tylotus* Zone is distinctly marked by an index species that appears in Bed 3 of the Mavrokolimbos section at the level 1.2 m above the base. This index species and associated forms occur at higher levels all along the Kannaviou Formation succession and within the basal part of the Kathikas Formation, where it encloses lenses of bentonite clays. Therefore, deposits of the *tylotus* zone have a considerably greater thickness than beds of the previous radiolarian zone. This may also indicate that the *pseudoconulus* zone is a condensed succession, or it may indicate that only its upper part is present in this section. The first interpretation is inconsistent with the monotonous lithology of the bentonite clays that are observed all over the Kannaviou Formation section, including its *tylotus* zone, where these deposits are thick and do not display the lithologic features indicative of decelerated sedimentation, rewashing, and other events that could be responsible for the formation of a condensed succession.

Radiolarians of the oceanic *tylotus* zone include some species that are characteristic of its lower part and became extinct during the Maestrichtian (Sanfilippo and Reidel, 1985). In the studied sections, these radiolarians are *Afens lirioides, Alievium gallowayi, Clathropyrgus titthium, Dictyomitra koslovae, Pseudoaulophacus floresensis*, and *P. parqueraensis*. They obviously indicate that only the lower part of the zone (upper Campanian–lower Maestrichtian) is represented in the described successions of the Kannaviou and Kathikas formations. This is consistent with the fact that, near the village of Kannaviou, the Kathikas Formation underlies upper Maestrichtian carbonate deposits. It is also interesting that, in Cyprus, the *Amphipyndax pseudoconulus* is distributed throughout the entire interval of the *tylotus* zone. In oceanic successions, this form became extinct in the very basal part of the *tylotus* zone, presumably before the Maestrichtian (Sanfilippo and Reidel, 1985). If this is true, and if *Amphipyndax pseudoconulus* has the same stratigraphic range in

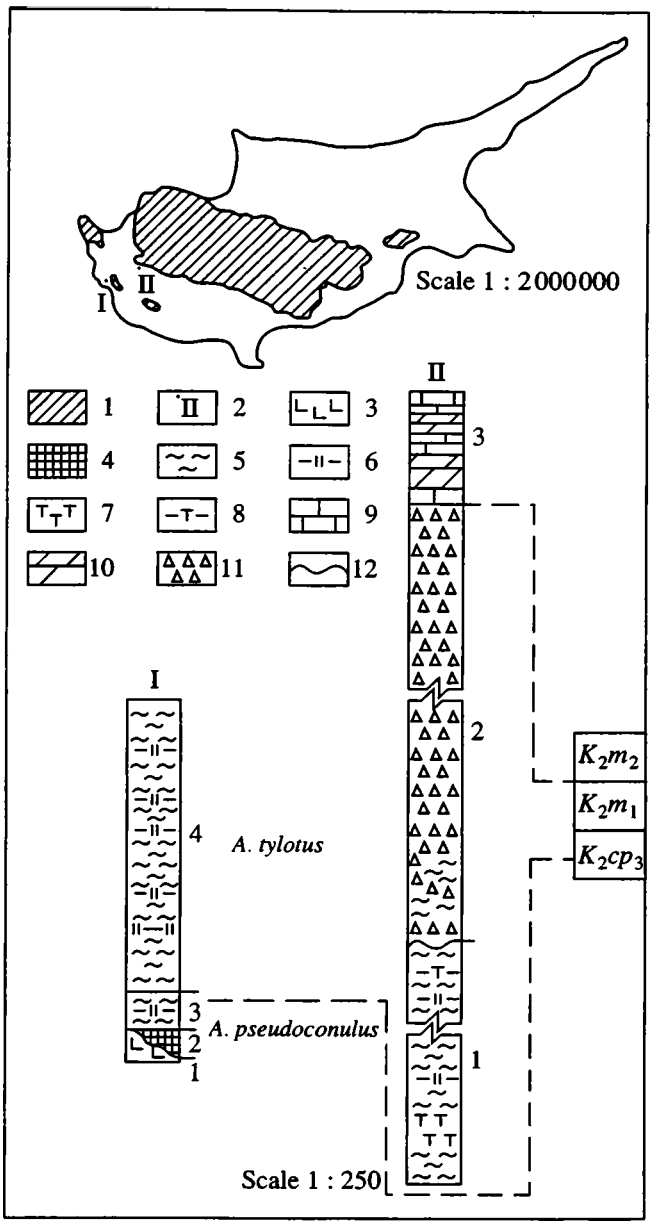
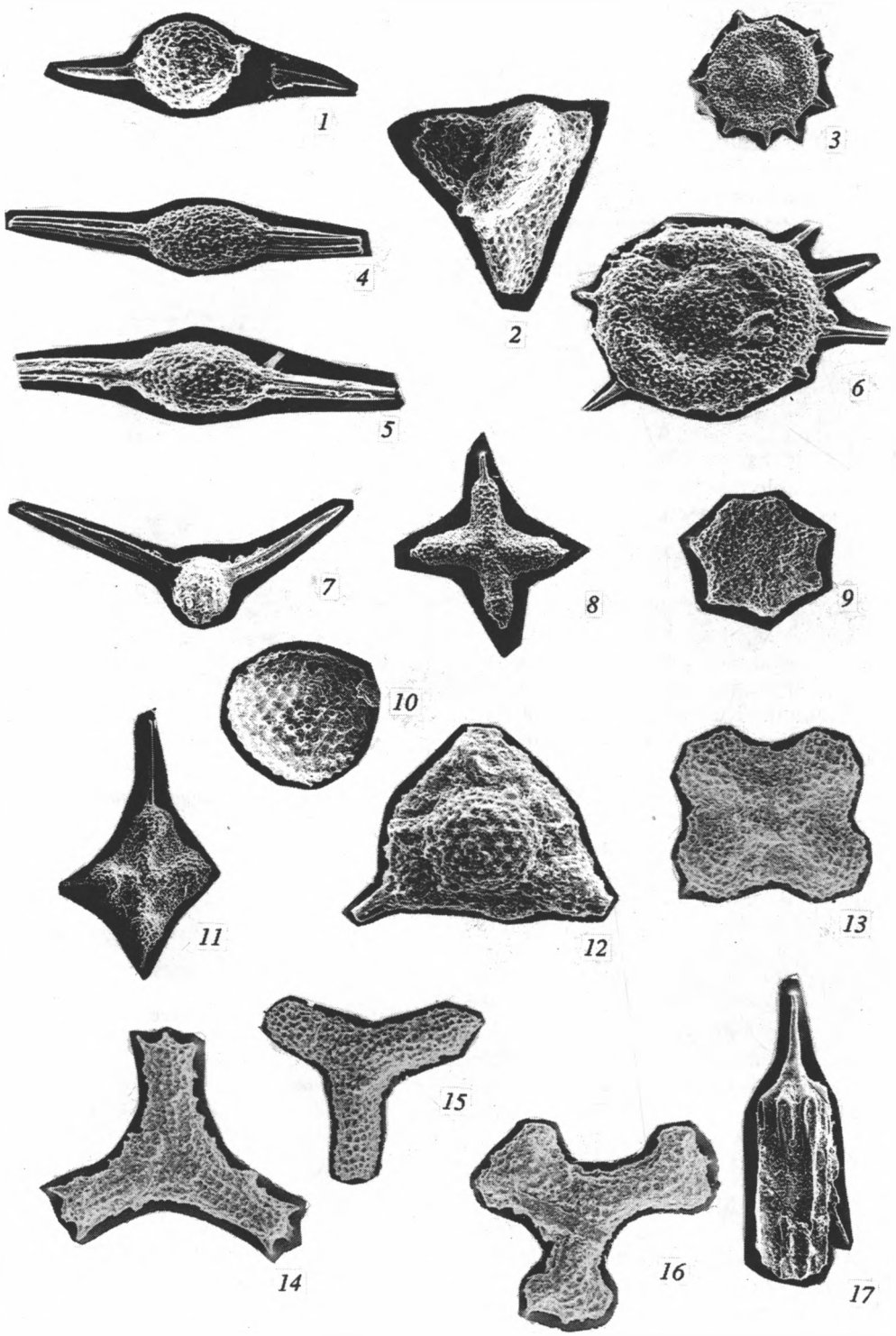


Fig. 1. Correlated reference sections of Upper Cretaceous deposits in southwestern Cyprus.

1 – ophiolites of the Troodos massif; 2 – section localities (I – Mavrokolimbos River, II – village of Kannaviou); 3 – basalts; 4 – metalliferous clays and umbers; 5 – bentonite clays; 6 – opokalike claystones; 7 – tuffs; 8 – tuffaceous siltstones; 9 – limestones; 10 – marls; 11 – mixtites; 12 – erosion, unconformity. Digits indicate bed numbers.

Plate III



Cyprus, then we have to admit that the deposits here that bear radiolarians of the *tylotus* zone correspond to a very narrow stratigraphic interval (upper Campanian?), despite the considerable thickness of the sediments.

It should be noted that the oceanic radiolarian zones of the Upper Cretaceous have significant stratigraphic ranges (Sanfilippo and Riedel, 1985). For example, the *Amphipyndax tylotus* Zone corresponds to an interval comprising the upper Campanian and the entire Maestrichtian. Nevertheless, inside this zone and the underlying *pseudoconulus* zones, several levels of the first occurrence and extinction of certain species that are interesting from a stratigraphic viewpoint were recognized (Sanfilippo and Riedel, 1985). If studied further, these levels would be useful in creating a more detailed scheme. Unfortunately, our successions revealed only a few of them, e.g., the extinction of *Theocampe urna* coincided with the basal level of the *tylotus* zone (Fig. 2). As for *Afens lirioides, Amphipyndax pseudoconulus, Clathropyrgus titthium*, and *Dictyomitra koslovae*, they are traceable everywhere in the radiolarian-bearing intervals of the studied sections in Cyprus, which, as was noted, could be a consequence of condensed sedimentation. Despite this, several levels are intriguing in the Mavrokolimbos section (Fig. 2), such as the first occurrences of *Crucella cachensis, Archaeospongoprunum bipartitum*, and associated forms in Bed 3 (6.5 m above the base). Still, to understand their stratigraphic significance we need additional information, first of all, data on the areal distribution and age compatibility of these forms.

The radiolarian assemblages distinguished in Cyprus are typical of Late Cretaceous tropical populations and are very similar to their groups that occur in oceanic regions of lower latitudes (Plates I - III). For example, 80 - 90% of their taxonomic composition are known in the Upper Cretaceous of California (Campbell and Clark, 1944; Foreman, 1968; Pessagno, 1976) and in the Campanian–Maestrichtian deposits of the Pacific Ocean (Riedel and Sanfilippo, 1974; Pessagno, 1975; Foreman, 1975; Sanfilippo and Riedel, 1985).

Plate III. *1* – *Stylosphaera pusilla* Campbell et Clark, Sample 36-10, *tylotus* zone; *2* – *Clathropyrgus titthium* Reidel et Sanfilippo, Sample 36-10, ×100, *tylotus* zone; *3* – *Spongodiscus* sp., Sample 36-2, ×75, *pseudoconulus* zone; *4* – Archaeospongoprunum tchamaensis Pessagno, Sample 36-10, *tylotus* zone; *5* – *A. salumi* Pessagno, Sample 36-10, *tylotus* zone; *6* – Same specimen as in *3*; *7* – *Acaeniotyle* ex gr. *diaphorogona* Foreman, Sample 36-10, *tylotus* zone; *8* – *Crucella messinae* Pessagno, Sample 36-10, ×75, *tylotus* zone; *9* – *Septinastrum dogeli* Gorbovetz, Sample 36-2, ×75, *pseudoconulus* zone; *10* – *Alievium* cf. *superbum* (Squinabol), Sample 36-8, ×75, *tylotus* zone; *11* – *Crucella plana* Pessagno, Sample 36-10, ×75, *tylotus* zone; *12* – *Pseudoaulophacus* ex gr. *riedeli* Pessagno, Sample 36-8, *tylotus* zone; *13* – *Crucella* aff. *robusta* Bragina, Sample 36-2, ×100, *pseudoconulus* zone; *14* – *Halesium* ex gr. *sexangulum* Pessagno, Sample 36-10, ×100, *tylotus* zone; *15* – *Rhopalastrum* (?) sp., Sample 36-8, ×100, *tylotus* zone; *16* – *Patulibracchium* sp., Sample 36-8, ×100, *tylotus* zone; *17* – *Afens lirioides* Reidel et Sanfilippo, Sample 36-10, *tylotus* zone.

In Cyprus, this group includes species of *Amphipyndax, Dictyomitra, Praeconocaryomma*, and *Pseudoaulophacus* genera as well as *Afens lirioides, Xitus asymbatos, Theocampe urna*, and some other forms. Nevertheless, there are also species that are absent or very rare in the aforementioned regions. These species are *Crucella latum, C. aster, Prunobracchium longum, Rhopalastrum tumidum*, and *R.* ex gr. *ingens*, known from the Upper Cretaceous deposits of the Russian platform (Lipman, 1952; Bragina, 1987) and the West Siberian plain (Kozlova and Gorbovetz, 1966). Presumably, these species can also be encountered in oceanic successions, because they are not endemics of the North Eurasian epicontinental basins and have a wider distribution. In the middle part of Bed 4 of the Mavrokolimbos section, some of these taxa (*Crucella latum, Prunobracchium longum*, and *Rhopalastrum* ex gr. *ingens*) occur within a narrow interval 6.5 - 8 m above the base and have not been distinguished outside of it (Fig. 2). All these forms are insignificant in abundance, and their appearance was probably related to some environmental changes. However, this remains to be proven lithologically.

The obtained data provide more refined age determinations for the Kannaviou and Kathikas formations. Formerly, the Kannaviou Formation, which also bears *Globotruncana*, was considered as either a Maestrichtian (Mantis, 1970; Ealey and Knox, 1975; Blome and Irwin, 1985) or Campanian unit (Swarbrick and Robertson, 1980). Due to the recognition of the *Amphipyndax pseudoconulus* and *Amphipyndax tylotus* zones of the radiolarian scale, we suggest that it corresponds to the stratigraphic range of the Campanian–early Maestrichtian, bearing in mind that *Afens lirioides, Amphipyndax pseudoconulus, Dictyomitra koslovae*, and other forms that became extinct during the Maestrichtian occur here until the top of the formation. Swarbrick and Robertson (1980) referred the Kathikas Formation to the Maestrichtian. However, in the lower part, it bears radiolarians of the *tylotus* zone, suggesting that this level is the late Campanian in age. Accordingly, this formation, which underlies upper Maestrichtian carbonates, can be considered as the late Campanian–early Maestrichtian in age.

PALEONTOLOGICAL DESCRIPTION

Class Radiolaria

Order Spumellaria Ehrenberg, 1847

Superfamily Spongodiscoidea Haeckel, 1862

Genus *Crucella* Pessagno, 1971

Crucella robusta Bragina, sp. nov.

Plate II, Figs. 1, 4

Crucella aster (Lipman): Nakaseko and Nishimura, 1982, Pl. 2, Figs. 9, 10.

Crucella? sp. A: Empson-Morin, 1984, Pl. 1, Figs. 4, 5.

Crucella sp.: Gorka, 1989, p. 332, Pl. 11, Fig. 5.

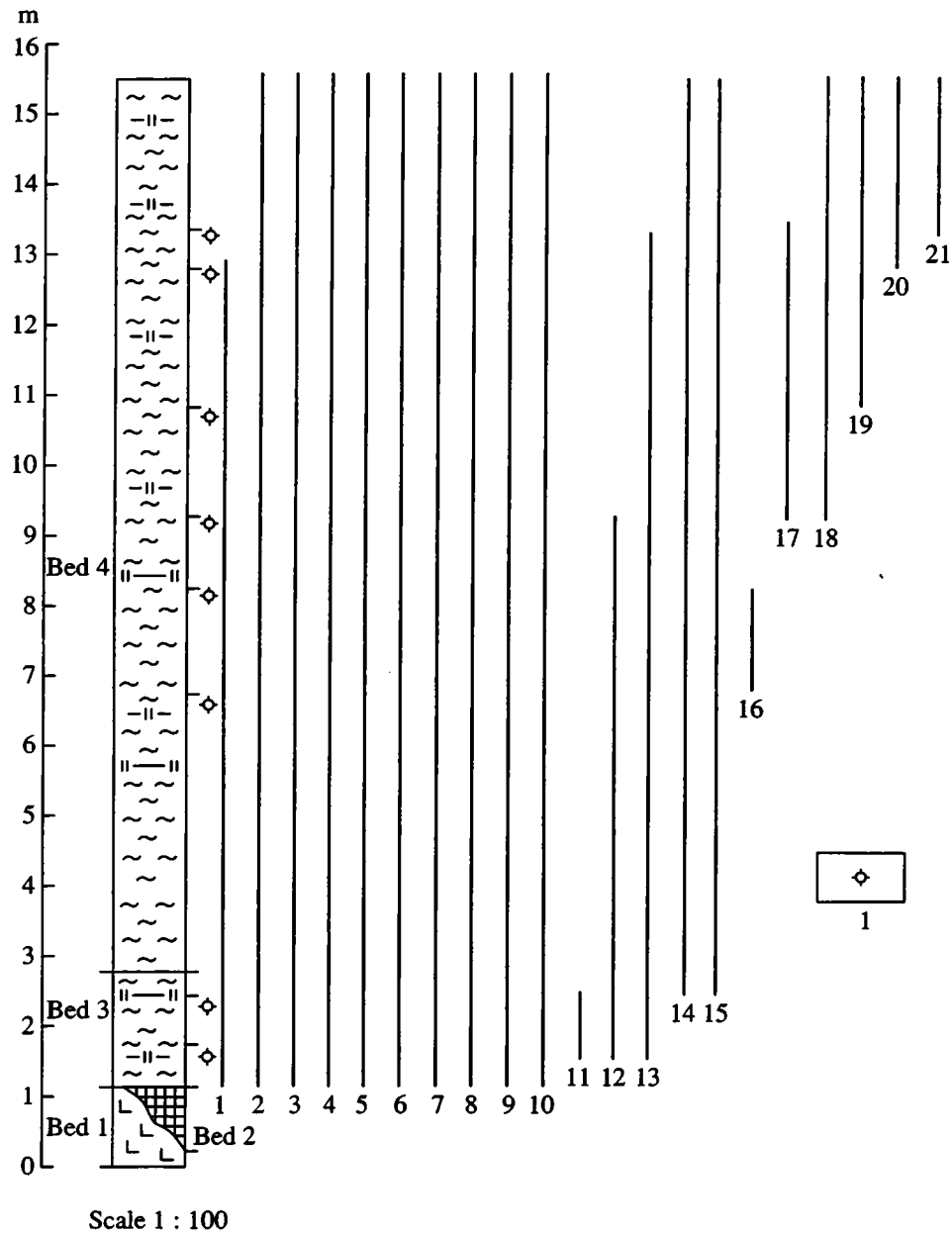


Fig. 2. Stratigraphic ranges of radiolarian species in the Mavrokolimbos section.

Species: 1 – *Dictyomitra duodecimcostata*; 2 – *Alievium gallowayi*; 3 – *Amphipyndax pseudoconulus*; 4 – *Afens lirioides*; 5 – *Dictyomitra koslovae*; 6 – *Dictyomitra formosa*; 7 – *Dictyomitra torquata*; 8 – *Patellula verteroensis*; 9 – *Thanarla veneta*; 10 – *Theocampe urna*; 11 – *Crucella messinae*; 12 – *Dictyomitra multicostata*; 13 – *Stichomitra cechena*; 14 – *Amphipyndax tylotus*; 15 – *Hemicryptocapsa conara*; 16 – *Phaseliforma laxa*; 17 – *Clathropyrgus titthium*; 18 – *Prunabracchium longum*; 19 – *Acaeniotyle* ex gr. *diaphorogona*; 20 – *Acidnomelos* (?) sp.; 21 – *Theocampe* ex gr. *ascalia*.

1 – levels of sampled radiolarians. Bed 1 - Bed 4 – numbers of beds.

Holotype. Geol. Inst. 4839/1. Southwestern Cyprus, the lower course of Mavrokolimbos River, late Campanian–Maestrichtian, Kannaviou Formation, Sample 90B36-8.

The name is from the Latin *robustus*, meaning an oak, strength.

Description. Crucella has massive arms that are abrupt at the ends and sometimes terminate with a short spine. The prominent tetragonal tallus has a depressed central part. The arching axial parts of the arms have, similar to the tallus, a pseudoaulophacoid structure. The distal parts of the arms are spongy, and they grow through various ontogenetic stages until they are coalescent with each other. The shell then has a tetragonal form, with the axial parts of the arms radiating toward its rounded angles.

Dimensions. The shell is 200 - 300 μm wide; the free-arm ends are up to 60 μm long; and the arm width is 60 - 100 μm.

Comparison. This form is differentiated from *Crucella cachensis* Pessagno by its very massive arms with abrupt ends. It differs from *Crucella aster* (Lipman) in the pseudoaulophacoid walls of the higher tallus and the arching axial parts of the arms.

Material. Several dozen well-preserved shells from two localities.

Distribution. The Upper Cretaceous (upper Campanian–Maestrichtian) of Cyprus (Kannaviou and Kathikas formations) and Upper Cretaceous oceanic successions (Reidel and Sanfilippo, 1985; Empson-Morin, 1984).

Reviewers K.I. Kuznetsova and I.A. Basov

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Age of the Natskhor and Shiraki Formations of Eastern and Central Georgia and the Problem of the Miocene–Pliocene Boundary: Paleomagnetism and Stratigraphy

G. Z. Gurarii*, I. A. Kudasheva*, V. M. Trubikhin*, A. G. Fein, and S. V. Shipunov***

** Institute of Geology, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia*
*** Shmidt Institute of Earth Physics, Russian Academy of Sciences, Bol'shaya Gruzinskaya ul. 10, Moscow, 123810 Russia*

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Abstract – Paleomagnetic and stratigraphic studies were performed on the terrigenous sedimentary complexes that occur in central and eastern Georgia between the Sarmatian and Akchagylian deposits. It was established that the complexes named here as the Shiraki and Natskhor formations are of the same age and, in the most complete sections, correspond to the Meotian and Pontian as well as the Azov Horizon deposits of the Black Sea basin. Correlation with the magnetostratigraphic scale showed that they correspond to the fifth and sixth paleomagnetic epochs as well as to the lower part of the Gilbert epoch, up to and including the Tvera event. If the Miocene–Pliocene boundary was assumed to be between the Meotian and Pontian, it seems quite difficult to trace it everywhere in the studied deposits due to the genetic unity and homogeneity of the sediments forming the sequence.

Key words: *Magnetostratigraphy, stratigraphy, Neogene, Paratethys, Ponto-Caspian, Messinian, Meotian, Pontian, Shiraki Formation, Natskhor Formation, boundary.*

INTRODUCTION

The terrigenous sedimentary complexes between the Sarmatian and Akchagylian formations, which are known in the literature as the Natskhor and Shiraki formations, Lengebiz sandstones, and other names, are widespread in the southern Paleocaspian regions. As a rule, these formations are barren of fossils, and their dating varies from the Late Sarmatian to the Middle Pliocene (Alizade *et al.*, 1972; Buleishvili, 1960). Nevertheless, the problems of age determination, mutual correlation, and comparison with similar deposits of other regions as well as with the magnetostratigraphic scale appear to be topical in light of the problem of the Miocene–Pliocene boundary (Nevesskaya *et al.*, 1984).

This article presents the results of paleomagnetic and stratigraphic studies on the Shiraki (Kakheti, eastern Georgia) and Natskhor (Kartli, central Georgia) formations.

METHODS OF INVESTIGATIONS

Three sections of the Shiraki Formation in the Iori River valley and a section of the Natskhor Formation on the left side of the Kura River were used for the paleomagnetic investigations (Fig. 1). These sections were chosen because they are thick, complete (Eldar and Caspi sections, Fig. 1, points 4 and 1), and bear fauna (Dzedvtakhevi section; Figs. 1 and 2). The Kotsakhuriskedi section (Figs. 1 - 3) is intermediate between the Eldar and Dzedzvtakhevi sections. Their thicknesses are 1100, 700, 400, and 200 m, respectively. Paleomagnetic lump samples were picked by hand at intervals of 10 - 20 m. After laboratory studies on the primary collection, resampling from the intriguing parts of the section was performed at reduced intervals (1 - 2 m). In a number of cases, the sampling intervals were reduced to obtain a more exact restriction of the paleomagnetic boundaries. As a result, 550 samples were gathered from the studied sections, 2.5 km in total thickness; 4 cubic samples with 2.0-cm edges were cut out from each sample. Simultaneously, we completed a detailed bed-by-bed lithological description of the sections.

Magnetomineralogical investigations were performed in the laboratory with the goal of determining the composition of the magnetic minerals responsible for the natural remanent magnetization (NRM). For this purpose, we studied the dependence of isothermic remanent magnetization (IRM) on the steady magnetic field (H) (up to the saturation state: SIRM), determined the remanent coercive force (Hcr), performed a thermomagnetic analysis of SIRM(T) and SIM(T), and determined the variations of SIRM and Hcr under a stepwise heating of the samples. In addition, we studied variations in magnetic susceptibility along the investigated

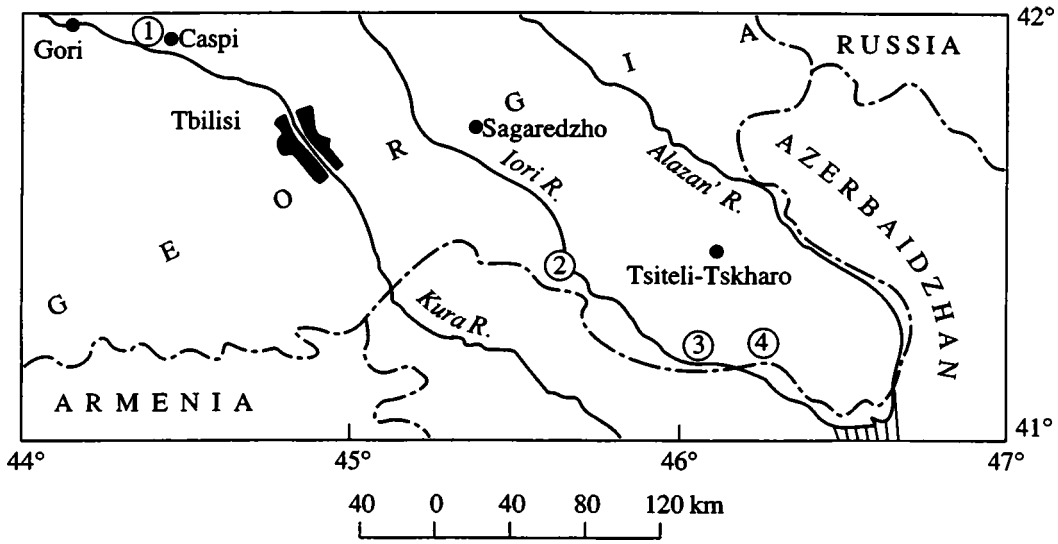

Fig. 1. Section localities.
1 – Caspi-1, 2 – Dzedzvtakhevi, 3 – Kotsakhuriskedi, 4 – Eldar-3.

sections and examined polished thin-sections. It should be noted that all of these investigations were used to study individual intervals of the Eldar section, which was adopted as the reference standard (especially for those parts of the section in which the fine structure of the geomagnetic field was studied). The extreme homogeneity of rocks forming other parts of this and additional sections suggest that they are compatible with the drawn conclusions.

To determine the NRM–ChRM characteristic component, which is the principal one in our investigations, we performed thermal cleaning in an oven placed behind a double screen made of μ-metal. To decide upon the temperature for cleaning, we used the data on the complete stepwise demagnetization of the leading collection. After analyzing the results of mass cleaning at $T = 250$ - 280°C, we performed a complete thermal demagnetization of the samples. However, the obtained results did not allow the NMR sign to be determined unambiguously. Two portions of each original sample were subjected to cleaning every time

RESULTS

Magnetic Minerals and the Nature of ChRM

The results of magnetomineralogical investigations allowed us to draw the following conclusions: (1) Magnetite, sometimes oxidized, is the basic ferromagnetic mineral. (2) The studied section is rather homogenous, relative to its ferromagnetics. These conclusions are confirmed by the data on the Hcr magnitude (18 - 30 mT) as well as by the study of the thin-sections. Here, the form of the magnetite grains and their relations with the host minerals suggest that the magnetite probably has a detrital origin (at least, those magnetite grains that are 10 μm or more in dimension). This is a possible explanation for the orientation of the main ancient NRM. At the same time, the presence of oxidized grains of magnetite (and, probably, hematite in some cases) suggests that ChRM in some intervals of the section (sections) reflects a complex magnetization, including the primary and secondary (corresponding to the period of oxidation) components. More comprehensive information will be published elsewhere (Gurarii and Kudasheva, in press).

Results of Thermal Cleaning

Figure 2 shows the results of the complete demagnetization of NRM. It is clear that the characteristic magnetization appears by heating the samples to 200 - 250°C. Before that point, the viscous magnetization is destroyed. ChRM identified in this manner can be considered as DRM, i.e., it shows the state of the magnetic field during the rock formation. Other samples were heated to $T = 250$ - 280°C. The data obtained are presented in Figs. 3 - 5. It is evident that in some cases the magnetization of individual samples shows an intermediate direction, which confirms the presence of secondary magnetizing components. The secondary magnetization obviously corresponds to the period of magnetite oxidation and was destroyed simultaneously with the primary component (their blocking temperatures overlap). In this regard, the ChRM direction of such samples can be interpreted only by taking into account all the data obtained on the adjacent sampling levels.

Results of Stratigraphic and Paleomagnetic Studies on the Sections

The most complete section of the Shiraki Formation, Eldar-3, is located on the left side of the Iori River in a lowland surrounding the Eldar steppe on the north-

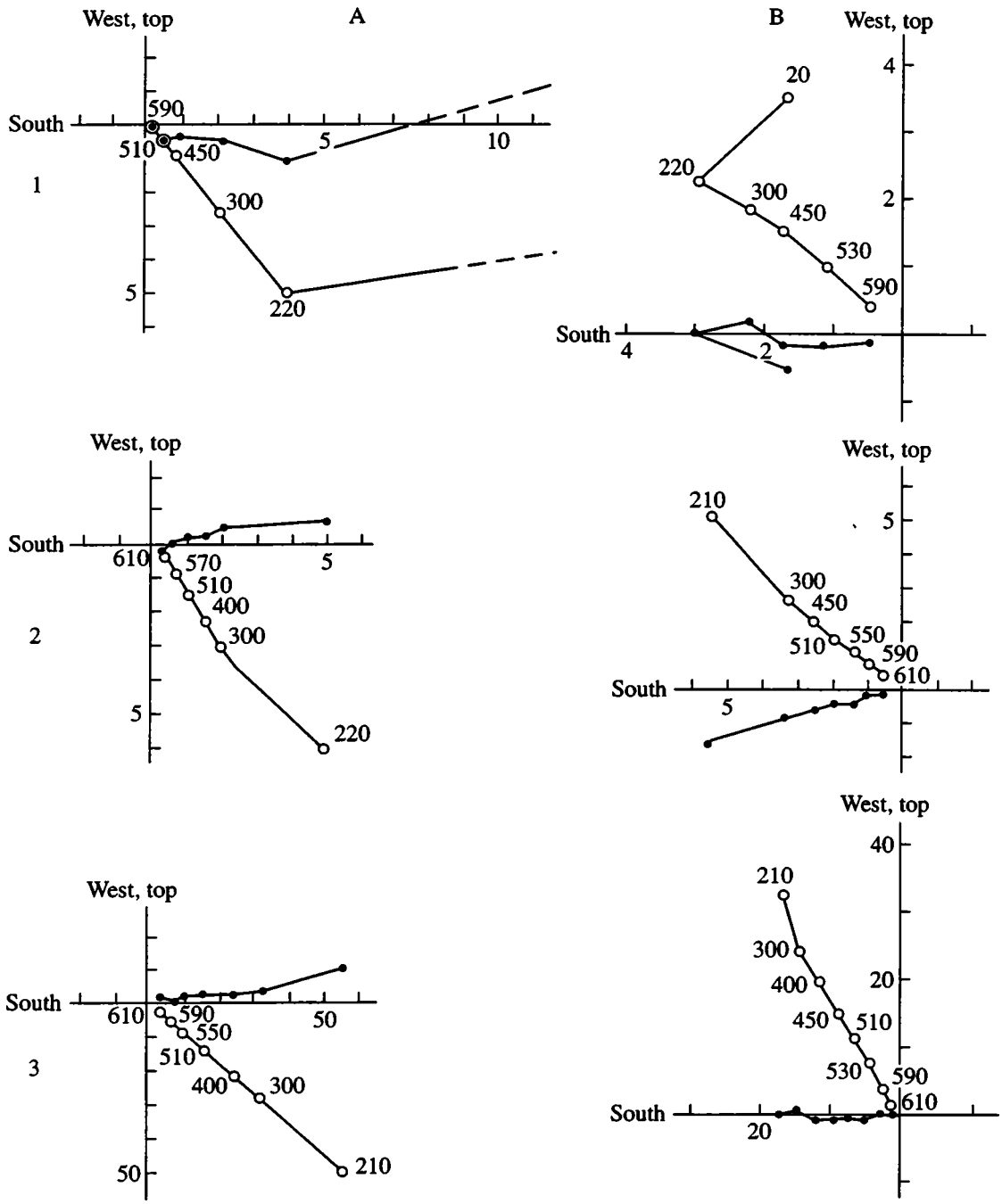

Fig. 2. Orthogonal-vector diagrams of complete demagnetization.
1 – Caspi-1, 2 – Dzedzvtakhevi, 3 – Eldar-3.
Open and closed circles designate projections on vertical and horizontal planes. A, B – for samples of normal and reversed polarity zones, respectively. All values are corrected for bed inclinations. Intensity of magnetization: 10^{-3} Am^{-1}.

western side. Here, a thick sequence of terrigenous deposits of the Shiraki Formation covers with an angular unconformity and erosion the speckled clays and sandstones of the Eldar Formation (the upper Sarmatian), which includes fauna. In the lower part of the section, the rocks have a dip azimuth of 355° and a dip angle of 32°, while the underlying rocks of the Eldar Formation have a dip azimuth of 334° and dip angles of 40° - 42°. The total thickness of the Shiraki Formation deposits in this section exceeds 1000 m. This is why the section, which represents a rather gentle monocline, is very extensive. The aforementioned parameters change along the section and gradually become more gentle (up to 22°) toward the top; the strike remains the same.

The Shiraki Formation deposits are covered with an angular unconformity and erosion by the Akchagyl conglomerates. The section is nearly barren of fauna. It was recovered only at a level of 500 m below the top, where poorly preserved imprints of unionids were found.

It seems unnecessary and irrational to provide in this article a complete bed-by-bed description of the 1-km-thick Shiraki Formation because of its monotonous lithology. The section consists of a rather homogenous succession of rhythmic members with an average thickness of 40 - 50 m. As an example, we provide a descripton of two rhythmic members in the lower and upper parts of the section.

At a level of about 770 m below the top and upwards the section, the following deposits are exposed in rhythmic member 4:

(1) Grey coarse- and medium-grained sandstones: 3 - 4 m.

(2) Greenish-brown siltstones with sand interbeds: 22 m.

(3) An interbed of grey coarse-grained compact sandstone: 0.5 m.

(4) Brownish-grey sandy siltstones: 7 - 8 m.

The total thickness of the member is about 35 m.

At a level of about 230 m below the top and upwards the section, rhythmic member 16 consists of the following deposits:

(1) Grey coarse-grained sandstones: 2 m.

(2) Variegated bluish- and yellowish-grey siltstones: 25 m.

(3) Grey, medium-grained sandstones with interbeds of yellowish-brown sandy siltstone: 4 m.

(4) Variegated bluish- and yellowish-grey siltstones: 35 m.

(5) Brownish-grey coarse-grained sandstones: 1 - 3 m.

(6) Grey, highly arenaceous siltstones (up to sandy loams) with interbeds of compact grey sandstones: 13 m.

The total thickness of this rhythmic member is about 80 m.

Such rhythmic members (hereafter referred to as "rhythms") form the entire section (Fig. 3). However, it also has some more general features. According to the content of sandy components, which was estimated by the proportions between the thicknesses of sandstones (conglomerates) and siltstones for each rhythm, there is a general increase in the content of coarse material upward the section. Its lower part incorporates mostly medium- and coarse-grained sandstones, while in the upper part, coarse-grained sandstones and conglomerates prevail along with clayey to highly arenaceous siltstones (including sandy loams). In addition, the content of sandy components varies along the section, dividing it into three parts. The minimal content of sandy components (about 5, 7, and 9%) is observed in the lowest part of the section (rhythms 8 and 16, respectively).

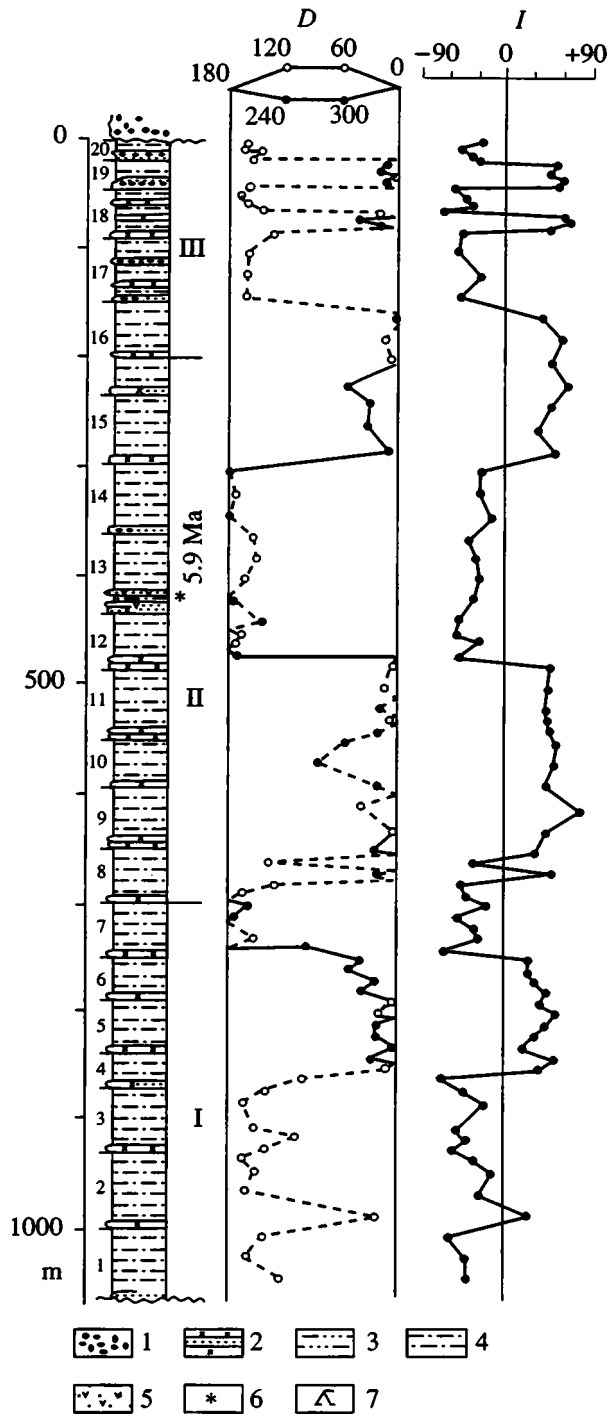

Fig. 3. Eldar-3 Section (Shiraki Formation).
1 – conglomerates and breccia, 2 – sands and sandstones, 3 – sandy siltstones, 4 – siltstones, 5 – ashes, 6 – track dating, 7 – locality of vertebrate fauna.
To the left are the numbers of rhythmic members, and to the right are the numbers of the sedimentary complexes.

The maximum content of sandy components (approximately 20, 25, and 35%) is in rhythms 5, 13, and 17, respectively. Thus, the sequence of the Shiraki Formation in the Eldar-3 section is divisible into three sedi-

mentary complexes, which seem to correspond to three stages of active tectonic movements in the region and, respectively, to three stages of activated sedimentation.

Among the features of the described section, it seems necessary to note, first, the appearance of brightly colored (red-brownish and bluish-grey) members of sandstones in the lower part of the middle complex and, second, the presence of an interbed of grey volcanic ash in the middle part (rhythm 13). This ash was dated using the track method (Chumakov *et al.*, 1992). Depending on different constants for the spontaneous fission of uranium, it is approximately 7.1 or 5.9 Ma old.

Paleomagnetic study of this section showed that the deposits that compose the upper 150 - 160 m (rhythms 17 - 20) have, in general, a reversed magnetization [Block A with two horizons (a double horizon) of normal polarity within the interval 25 - 80 m below the top of the section]. Below this block there is another zone about 500 m thick (rhythms 9 - 16), which may be considered as a zone of normal polarity (Block B with a thick horizon of reversely magnetized rocks within the interval 305 - 475 m below the top of the section). Finally, a zone of mainly reversed magnetization corresponds to Block C (thickness of 400 m, rhythms 1 - 8) with a horizon of normal polarity within the interval about 750 - 850 m below the top of the section.

The next section is located 20 km to the west, on the southern slope of the Kotsakhuriskedi Ridge, the left side of the Iori River. Here, the Akchagylian conglomerates cover a member of loam containing unionids of the lower Poratian. This loam covers with unconformity the deposits of the Shiraki Formation. The total thickness of the studied section is about 200 m. There are no outcrops below this area, because the Shiraki deposits are covered by the alluvium of the Iori River. The dip azimuth of the former is 0°, and the dip angle is 20°. In general, the Shiraki deposits here are the same as those in the previous section; the only difference is that the content of sandy components is slightly lower (about 13% for the section as a whole). Paleomagnetic studies demonstrated that this section corresponds to a zone of normal polarity and is probably analogous to Block B of the preceding section. It comprises a horizon of reversed polarity about 50 m thick within the interval 80 - 130 m below the top of the section (Fig. 4).

The Dzedzvtakhevi section is located at a distance of about 40 km northwestward, on the left side of the Iori River, where a large occurrence of vertebrate remains was discovered (Vekua *et al.*, 1988). Here, as in the Eldar section, the Shiraki deposits unconformably rest on the upper Sarmatian variegated clays of the Eldar Formation, which are covered with an unconformity and erosion by the Akchagylian conglomerates. Unfortunately, the lower deposits of the section are fragmentarily exposed and, therefore, have not been studied. An upper sequence of sandstones and siltstones about 400 m thick is exposed. These deposits form a gentle monocline with a dip azimuth of 22° and a dip angle of 21°. As in the Eldar section, the deposits compose successive rhythmic members, and the coarse material content increases upward (Fig. 4). The results of paleomagnetic studies allow the correlation of both sections, and, below, we describe of a rhythm synchronous with an equivalent member of the Eldar section (rhythm 16 within the interval 150 - 230 m below the top of the section).

At the interval about 40 - 90 m below the top of the Dzedzvtakhevi section (rhythm 8), the following deposits are exposed (upward the section):

(1) Massive coarse-grained yellowish-grey sandstones with bun-shaped jointing: 2 - 3 m.

(2) Yellowish-grey sandy siltstones: 3 m.

(3) Grey coarse-grained sands and sandstones: 2 m.

(4) Light-grey siltstones: 5 - 6 m.

(5) Yellowish viscous sandy loams with abundant pelitic material and numerous remains of vertebrates: 2 - 2.5 m.

(6) Variegated bluish-grey and yellowish-grey siltstones: 17 - 18 m.

(7) An interbedding of grey sandy siltstones and yellowish-grey coarse- and medium-grained sands: 14 - 15 m.

The total thickness of the rhythmic member is about 50 m.

The content of sandy components is 10% in the lower part; further, it increases to 20% in rhythms 7 and 8 and then reduces again to 13 - 14%. Paleomagnetic studies demonstrated that the upper 360 m of the section (rhythms 2 - 9) generally have normal polarity, though in the interval 120 - 180 m below the top of the section, there is a horizon with reversed polarity. Apparently, we are dealing here with the same Block B that was distinguished in two other sections. This assumption agrees well with the character of variation in the content of sandy components throughout the section. The lower part of the section is reversely magnetized (a part of Block C).

On the basis of vertebrate fauna (predominantly Mammalia), the complex of Shiraki deposits is dated by the Meotian and Pontian (Vekua *et al.*, 1988).

The Natskhor Formation section was studied on the left side of the Kura River, 5 - 6 km west of the town of Caspi. Here, a sequence of terrigenous deposits that are lithologically similar to those of the Shiraki Formation covers with angular unconformity and erosion the middle Sarmatian brown clays, which include fauna. In the Sarmatian clays, the dip asimuth is 20°, and the dip angle is 52°; the same parameters for the overlying deposits of the Natskhor Formation are 20° and 45°, respectively. The total thickness of the Natskhor Formation is about 700 m. The Natskhor deposits, which are barren of fauna, are covered by a sequence of conglomerates, sandy loams, and siltstones with a low degree of consolidation and dislocation, which probably testify to their late Pliocene age. The Natskhor Formation in the Caspi section is represented

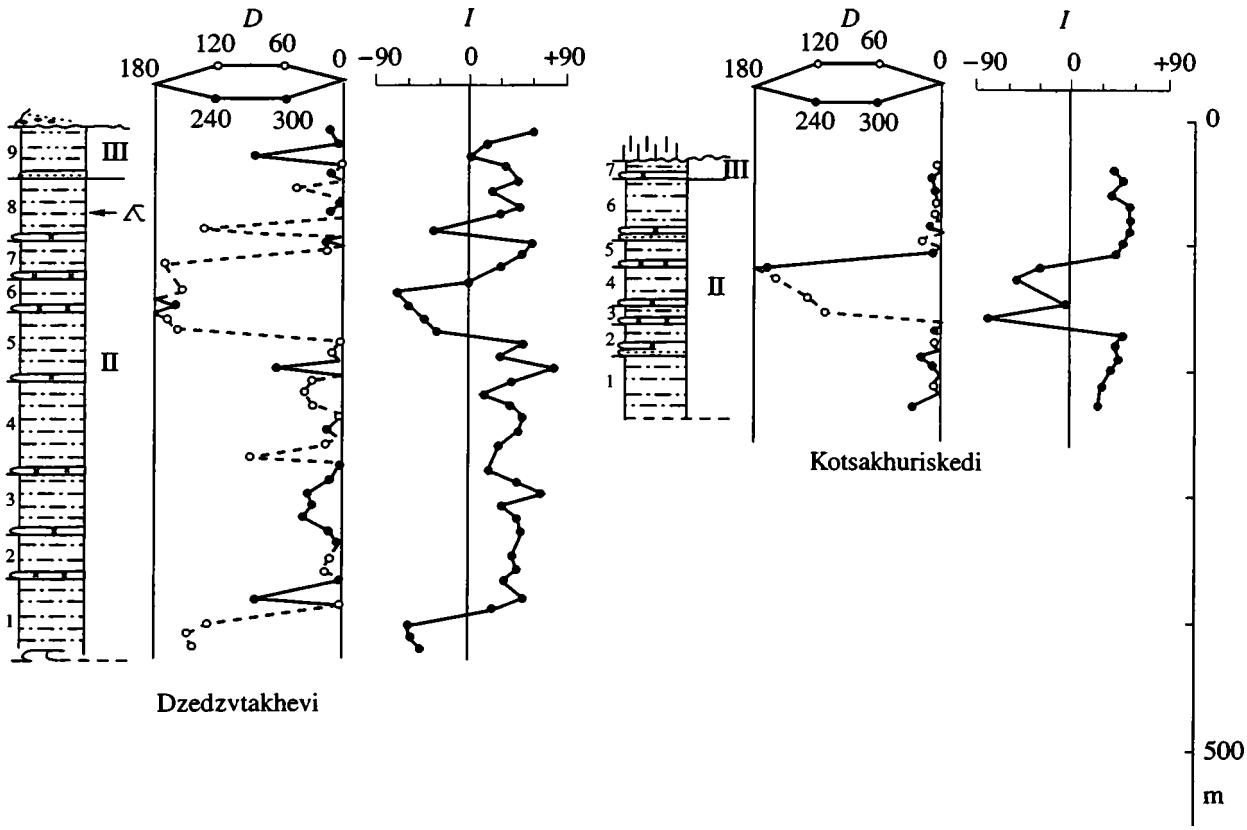

Fig. 4. Dzedzvtakhevi and Kotsakhuriskedi sections (Shiraki Formation). See Fig. 3 for key to lithology.

by a rhythmic sequence of coarse grey sands, brownish- and greenish-grey siltstones, and upper highly arenaceous siltstones and conglomerates (Fig. 5). The content of coarse terrigenous material, as in the previously described outcrops, increases up the section. As an example, we describe a rhythmic member in the upper part of the section.

The interval 35 - 70 m below the top of the section (rhythm 13) is composed of the following deposits (from the base upward):

(1) Conglomerates and grey coarse-grained sandstones: 5 m.

(2) Brownish-grey siltstones and sandy siltstones with sandstone interbeds: 20 m.

(3) Grey compact coarse- and medium-grained sandstones: 4 m.

(4) Brownish-grey highly arenaceous siltstones: 5 - 6 m.

The total thickness of the rhythmic member is about 35 m.

In addition, there are general variations in the distribution of coarse terrigenous material throughout the section. A relatively low percentage of sandstones is characteristic of the lower part and rhythms 8 and 13 (approximately 15, 25, and 27%, respectively). The maximum content of sandy components is noted in rhythms 7, 9, and 10 as well as in the uppermost part of the section, where it amounts to 30 - 40%.

Paleomagnetic studies showed that deposits composing about 50 m of the upper part of the section (rhythms 13 - 15) are reversely magnetized. They overlie a thick sequence of normally magnetized deposits (the interval 50 - 330 m below the top, rhythms 13 - 7) that contains a horizon with reversed polarity in the interval 110 - 150 m. Still lower, there is another thick sequence of reversely magnetized deposits (rhythms 1 - 7) with a horizon of normal polarity in the interval about 500 - 590 m below the top of the sequence.

As in the Eldar section, we denote these three large intervals with different polarity as blocks A, B, and C, respectively.

DISCUSSION

Correlation of the Shiraki and Natskhor Formations

Some authors (Buleishvili, 1960; and others) attribute the brown siltstones (red and variegated clays) that underlie the above section in the Gori–Caspi region to the Natskhor Formation as well. Moreover, they suggest that the brown siltstones conformably overlie the middle Sarmatian carbonate deposits and have a gradual transition to both the latter and the upper Natskhor Formation. Since the interbeds in the brown siltstones

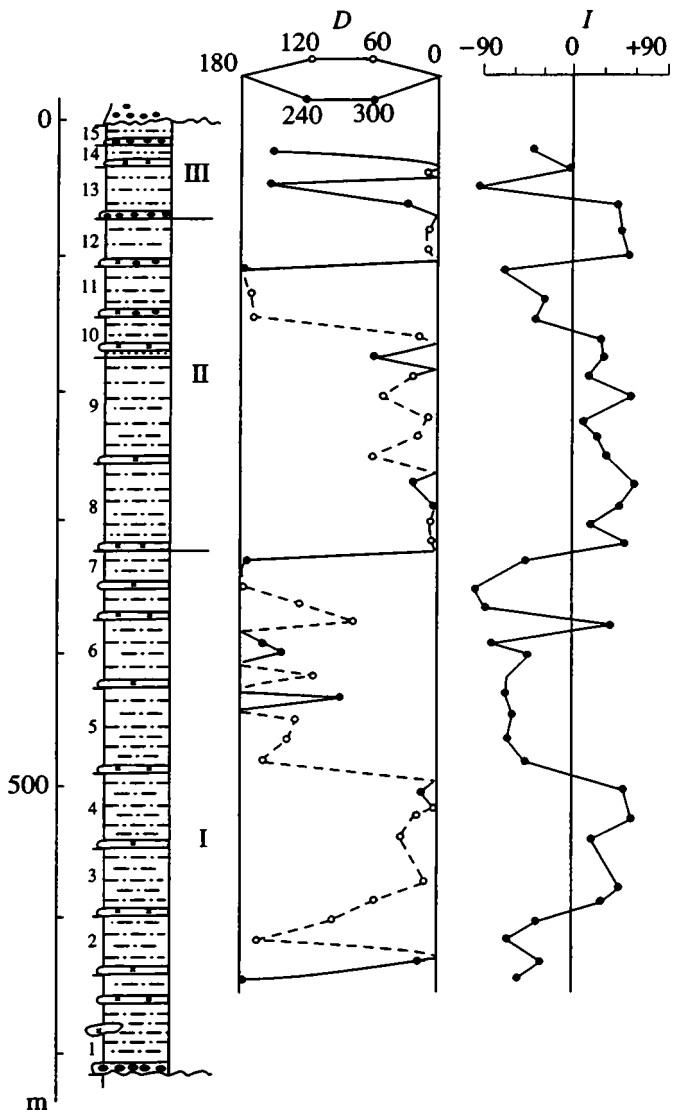

Fig. 5. Caspi-1 section (Natskhor Formation). See Fig. 3 for key to lithology.

contain middle Sarmatian fauna, the actual Natskhor deposits are considered as upper Sarmatian. However, as noted above, the Natskhor deposits cover the brown siltstones with an angular unconformity and deep erosion. The significance of the erosion is evident from the following observations. The thickness of the brown siltstones in the above-described outcrop does not exceed 50 - 60 m, and in the upper half of the sequence they incorporate a carbonate interbed with middle Sarmatian fauna. In the Nadarbazevi gully, 2 - 3 km westward, the thickness of the brown siltstones is about 200 m and, without any change in facies, they contain four carbonate interbeds.

Thus, the Natskhor and Shiraki formations are separated from the underlying Sarmatian deposits by a structural unconformity and are covered by upper Pliocene sediments. Figures 3 - 5 show that both formations are identical in their paleomagnetic characteristics and conditions of sedimentation. The variations in the distribution of sandy components in the Natskhor Formation section are similar to those in the most complete Eldar section of the Shiraki Formation. These variations allow the three correlative sedimentary complexes that compose both formations to be distinguished. This suggests that the Natskhor deposits in the Gori–Caspi region are not the late Sarmatian in age, but are analogous to the Shiraki Formation in Kakheti.

Age and Stratigraphic Range of the Shiraki and Natskhor Formations

Most of investigators date the Shiraki Formation at the Meotian–Pontian interval.The data obtained on the studied section favor this opinion. The track dating of ash beads in the Eldar section showed that they are 7.14 or 5.96 Ma old. For stratigraphic correlation, the uncertainty in the uranium decay constant is not very important. For example, in the Adzhiveli section (eastern Azerbaidzhan), which is characterized by fauna, the ashes near the boundary between the Meotian and Pontian are, according to paleomagnetic data, at a somewhat higher level than those in the Eldar section, where they were dated as 7.07 or 5.9 Ma (Chumakov *et al.*, 1992). The mammalian fauna from the Dzedzvtakhevi section, according to paleomagnetic data, occurs approximately at the level of dated ashes from the Adzhiveli section and is similar to the fauna of the well-known Bazaleti and Tudorovo localities, which are considered as the early Pontian–late Meotian in age (Vekua and Trubikhin, 1988).

Thus, the age of the Shiraki and, apparently, the Natskhor formations is not in question, but their stratigraphic range is another problem. In this case, the paleomagnetic results may be of decisive importance. Paleomagnetic investigations of the upper Neogene marine deposits with fauna (Black Sea basin) showed that the lower part of the Meotian is, in general, reversely magnetized in contrast to the upper Meotian, which has normal polarity. In the middle parts, both units have horizons with opposite polarity. The Pontian rocks usually have reversed magnetization, and the overlying Azov Horizon mainly corresponds to the interval (sometimes double) of normal polarity (Trubikhin, 1989). As for the Eldar section, which is the most complete of the studied sections, its lower part is represented by a thick zone of reversed polarity: Block C, which includes a subordinate horizon of normal polarity. Above, there is a thick zone of dominantly normal polarity with a horizon of reversely magnetized rocks: Block B. Almost all of this interval corresponds to the Meotian, because the track datings and the mammalian fauna from the upper part of Block B of the nearby section correspond to the uppermost part of the Meotian, adjacent to the boundary with the Pontian. Still higher, there is a zone of mainly reversed polarity: Block A, with two (double?) horizons of normal polarity. As noted above, such pat-

terns are characteristic of the Pontian–Azov Horizon interval.

Thus, in the most complete sections of the Shiraki Formation, its stratigraphic range corresponds to the joint interval of the Meotian, the Pontian, and the Azov Horizon. In two other sections of the Shiraki Formation, we observed similar magnetic patterns, excluding the upper zone of reversed polarity (Block A), which was probably destroyed by the pre-Akchagylian erosion. It is partially eroded in the section of the Natskhor Formation too. Thus, the upper boundary of both formations is distinct and has a structural character. The lower boundary of the Shiraki and Natskhor formations in the Eldar and Caspi sections is also structural in origin (angular unconformity and deep erosion). The deposits of the Shiraki and Natskhor formations were accumulated in the littoral part of a shallow basin of low salinity subjected to permanent, isostatically compensated downwarping, which was responsible for the facies character and the immense thickness of accumulated sediments. These depositional conditions remained unchanged throughout the Meotian and Pontian.

Thus, we are dealing here with a Meotian–Pontian structural and stratigraphic complex, which is separated by distinct structural and stratigraphic boundaries from the over- and underlying complexes of eastern and central Georgia.

The Miocene–Pliocene Boundary

Traditionally, the boundary between the Miocene and Pliocene in the eastern Paratethys was drawn at the base of the Pontian (Kolesnikov, 1940). However, as noted above, the definition of this boundary in the concrete sections that characterize the distribution area of terrigenous deposits of the Shiraki and Natskhor formations and their equivalents is, undoubtedly, rather difficult. The same situation is characteristic of sections containing marine deposits with fauna. In the Caspian basin (Adzhiveli section), and even in the stratotypical region of the Black Sea basin (Zheleznyi Rog section of Kerch and Taman'), the boundary between the Pontian and Meotian is free of remarkable features such as facial changes. As a result, if there is no continuous succession of mollusk fauna, the boundary between the Meotian and Pontian can be determined only tentatively. Nevertheless, the upper and lower boundaries of a unit coupling Meotian and Pontian (plus Azov Horizon) deposits is rather distinct. The examples are the unconformity between the Kamyshburun Beds and the Pontian and Azov Horizon deposits, similar relations between the productive formation and the Pontian of the Caspian basin, and the unconformable boundary of the Meotian deposits overlying the diatom formation of Kobystan. The unconformity and the bryozoan bioherms at the boundary between the Sarmatian and Meotian in the Kerch–Taman' region (regardless of the stage that these bioherms refer to) are also significant. Using these observations, some investigators have placed the Miocene–Pliocene boundary at the base of the Meotian (Zhizhchenko, 1958) or the Kimmerian (s. str.), i.e., the Kamyshburun Horizon (Semenenko, 1980).

Position of the Meotian–Pontian Interval in the Magnetochronological Scale and Its Analogues in the Mediterranean

Geological events related to boundaries such as the one between the Miocene and Pliocene may undoubtedly be of extraregional rank. This is why it is not only desirable, but necessary to correlate them with both the magnetostratigraphic (MSS) and magnetochronological (MChS) scales in order to reveal wide interregional relations. There are two principal viewpoints on the correlation of the Meotian–Pontian interval with the magnetochronological scale and the Mediterranean deposits of the same age.

According to the first viewpoint (Alexeeva *et al.*, 1983; Trubikhin, 1989), the Meotian corresponds, in general, to the fifth and sixth paleomagnetic epochs, while the Pontian, together with the Azov Horizon, corresponds to the lower part of the Gilbert epoch, up to and including the Tvera event. Thus, the entire Meotian–Pontian stage is comparable with the Messinian of the Mediterranean scale (variant 1).

According to the second viewpoint, the Meotian corresponds to the seventh and eighth paleomagnetic epoches, the Pontian corresponds to the sixth epoch, and the Azov Horizon corresponds to the fifth epoch together with the initial Gilbert epoch, including the Tvera event. A significant part of the Azov Horizon corresponds to a hiatus that lasted for about 1 m.y. and was typical of the entire Paratethys (Semenenko and Pevzner, 1979). In this case, the Meotian was correlated with the upper part of the Tortonian, and the Pontian was correlated with the upper part of the Messinian and the lower part of the Zanclian of the Mediterranean province (variant 2). Findings of nannoplankton in the Meotian and Pontian deposits of the Kerch–Taman' region served as the basis for this variant (Semenenko, 1980). However, the interpretation of these findings was, from the very beginning, ambiguous, and recently, the presence of autochtonous zonal forms in these deposits was called into question. As for the Azov Horizon, the hiatus that was proclaimed by the authors of the second variant seems to be far-fetched because of the obvious lithologic and faunal unity of these deposits, which represents, in our opinion, Pontian regressive facies. Although there is a number of track datings, they can be used to substantiate both the variants, depending on the choice of decay constants.

Thus, neither mollusk fauna nor plankton and track dating can serve as reliable stratigraphic markers. However, the considered interval is extremely rich in remains of vertebrates, both in the Paratethys and in the Mediterranean. As noted above, the mammalian fauna discovered in our Dzedzvtakhevi section is similar to

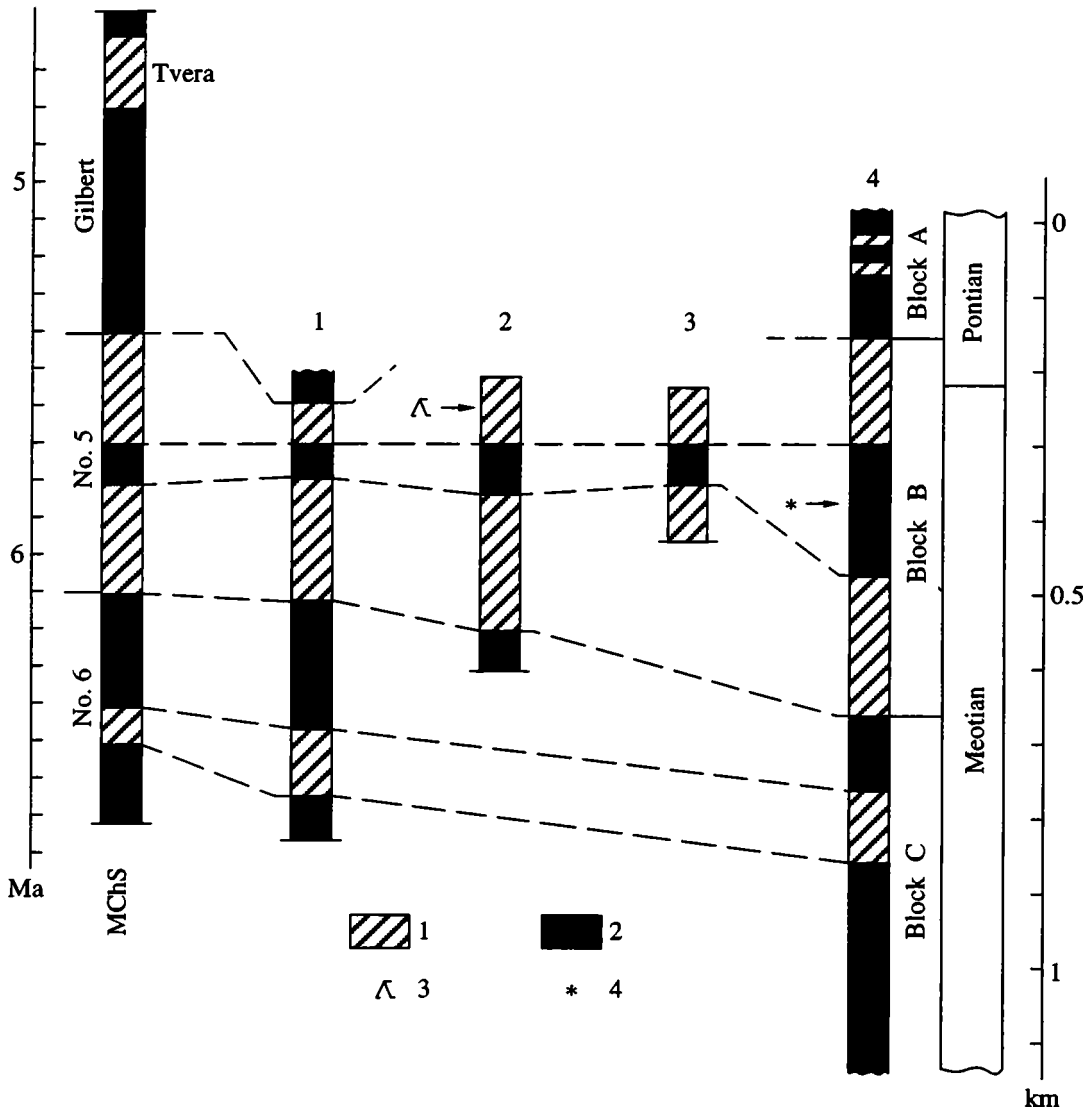

Fig. 6. Sections correlated versus magnetochronological scale (See Fig. 1 for the numbers of sections).
1 – normal polarity, 2 – reversed polarity, 3 – locality of vertebrate remains, 4 – track datings.
A, B, and C – magnetostratigraphic blocks.

that from the Tudorovo locality and occurs in the uppermost part of the Meotian interval. Remains from the Tudorovo are generally correlated with the Samos-5 level (Gabunia, 1981). This locality was also studied from a paleomagnetic viewpoint and was dated using the K-Ar method. The sampled interval with the Samos 1 - 5 fauna groups corresponds to the Messinian, and the Samos-5 level is near the episode of reverse polarity in the fifth epoch (Sevken Sen and Valet, 1986). Recently, the fauna of small mammalia from the marine Pontian deposits of the Odessa and Bolgrad (Vinogradovka) localities was studied. The obtained results show that we are dealing either with the uppermost Turolian or with the Ruscinian (Topachevskii *et al.*, 1988).

Thus, the reversely magnetized Pontian rocks of the lectostratotype in no way correspond to the sixth paleomagnetic epoch, and we have to correlate them with the lower part of the Gilbert zone. There are other reliable reference levels as well; however, in our opinion, the above-mentioned one seems to imply that the first variant correlating the Meotian–Pontian interval with the magnetochronological scale is more favorable.

Accordingly, we may conclude that the Meotian corresponds to the fifth and sixth epochs, and that the Pontian with the Azov Horizon correlates to the lower part of the Gilbert epoch, including the Tvera event. Thus, the interval as a whole correlates with the Messinian (Fig. 6).

At present, the Messinian neostratotype is composed of two supplementary sections of the Kapodarso and Paskuaziya localities (Ryan *et al.*, 1974). Here, the fol-

lowing succession is observed: lower marls, diatomites, and evaporites followed by upper marls (Lago-Mare) and evaporites. The base of the Lago-Mare formation correlates with the level of 5.4 Ma (Cita and McKenzie, 1986). This means that the lower sedimentary complex generally corresponds to the Meotian, and the upper one corresponds to the Pontian of the Paratethys. The upper and lower boundaries of the Messinian are not definitively identified, but the interval includes the fifth and sixth epochs as well as the lower part of the Gilbert epoch (Burckle, 1978). Thus, the Messinian (s.l.) is approximately of the same stratigraphic range and seems to have the same position in the magnetochronological scale as the Meotian–Pontian interval of the Paratethys.

CONCLUSION

(1) The results of our study on the Shiraki and Natskhor formations in eastern and central Georgia confirm the identity of their structural position and their lithologic and paleomagnetic characteristics and permit them to be dated at the same age.

(2) This terrigenous sedimentary complex is the Meotian–Pontian in age. The deposits of the Shiraki (Natskhor) Formation on the whole, along with the marine Meotian and Pontian (with the Azov Horizon) deposits, correspond to the fifth and sixth paleomagnetic epochs together with the lower part of the Gilbert epoch, including the Tvera event.

(3) These deposits are separated from the under- and overlying sediments by structural boundaries. Within these deposits, the boundaries between the Meotian and the Pontian with the Azov Horizon are distinguishable rather tentatively.

(4) In the Mediterranean, the Messinian (s.l.) sedimentary complex, which includes lower marls, diatomites, and evaporites, corresponds to these deposits.

(5) In the eastern Paratethys, the boundary between the Miocene and Pliocene corresponds to the boundary between the Meotian and Pontian. Nevertheless, the study of the barren terrigenous sequences and the marine sequences with fauna of this age shows that this boundary is not associated with any remarkable geological events and, in some cases, is conventional. In the Mediterranean, it may also run within the Messinian sedimentary complex. In practice, it is generally accepted as the boundary between the Messinian and Zanclian, which corresponds to the Miocene–Pliocene boundary. In the Paratethys it is, obviously, the boundary between the Azov Horizon and the Kimmerian s. str., i.e., the Kamyshburun Horizon. Some investigators prefer to place the Miocene–Pliocene boundary at the base of the Messinian or the Meotian, respectively, in both the Mediterranean and the Paratethys. Both of these versions seem more preferable than the idea of placing this boundary within the Meotian–Pontian interval.

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The Age of Rodent Fauna (Rodentia, Mammalia) from the Buran Formation of the Zaisan Depression, Eastern Kazakhstan

N. S. Shevyreva

Paleontological Institute, Russian Academy of Sciences, Profsoyuznaya ul. 123, Moscow, 117321 Russia

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Abstract – The Buran Formation is a member of continental Tertiary sediments located in the Zaisan depression. After screen-washing, sediments of this formation, which outcrops at the foot of the Manrak Range, yield diverse remains of small mammals, including rodents. The remains are represented by fragmental mandibles and individual teeth. A preliminary study of the materials shows that the rodents are represented by the families Castoridae (*Agnotocastor devius* sp. nov., *Steneofiber kazakhstanicus*, *S. zaysanensis* sp. nov., *S. shevyrevae* sp. nov., and *S.* aff. *shevyrevae*), *Cylindrodontidae* (*Cyclomylus mashkovae* sp. nov. and *Beatomus gloriadei* sp. nov.), and Cricetidae (*Cricetops collator* sp. nov. and *Leidymys asibaevi* sp. nov.) and by the superfamily Ctenodactyloidea (*Woodomys dimetron* sp. nov., *Muratkhanomys velivolus* gen. et sp. nov., *M. kulgayniae* sp. nov., *M. djanarae* sp. nov., *Roborovskia collega* gen. et sp. nov., *Tataromys raeda* sp. nov., and *T. boreas* sp. nov.). Analysis of accompanying groups of insectivores, bats, lagomorphs, fishes, turtles, and other vertebrates together with molluscs, ostracodes, charophytes, and macrofloral remains shows that the Buran faunal assemblage is not completely identical to the Oligocene faunas of Kazakhstan, Mongolia, and China. The Buran Formation cannot be considered as a strict analogue of the Chilikty Formation and other synchronous deposits of Mongolia and China. However, all faunas characterizing the commencement of the second half of the early Oligocene in Kazakhstan are more similar to the Buran fauna than to faunas of Central Asia. In comparison with European faunas, the stratigraphic level of the Buran fauna is higher than the levels of the Hoogbutzel and Ronzon zones and lower than those of the Bohningen and Antoingt zones. The age of the Buran Formation is considered here to correspond to the earliest middle Oligocene of older scales, or, according to the modern scale, the beginning of the second half of the early Oligocene.

Keywords: *Buran Formation, beginning of the second half of Early Oligocene, rodents, vertebrate groups, invertebrates, flora.*

The search for and collection of rodent remains are particularly significant for intensifying attempts to obtain detailed subdivisions of continental deposits on the basis of small mammal fossils (insectivores, lagomorphs, rodents, and others) all over the world. Rodents have a rapid change of generations, high degree of evolutionary flexibility, and closer links with substratum and vegetation than large mammals. Due to these biological features, rodents are key assemblages in the biozonal subdivision of deposits.

In comparison with other parts of the world, Asia is a poorly known region with respect to small mammal assemblages. First and foremost, these faunas are insufficiently studied in the Tertiary deposits of Kazakhstan, Central Asia, Indostan, the Near East, and southeastern Asia.

Faunas hitherto known or discovered during the past two decades are rare, separated by huge stratigraphic hiatuses, and geographically isolated. Given the present state of our knowledge about the Cenozoic small mammals of Asia, and considering that paleontological studies on the Paleogene have been irregular, even in the best-studied regions such as Mongolia, prospects are gloomy for creating a detailed biozonal scheme of stratigraphic correlations, similar to the one elaborated for Europe, in this region (Brunet and Vianey-Liaud, 1987; Crochet *et al.*, 1975; Mein, 1976).

These concerns are not simply related to Asia as a whole and adjacent land regions. Even in much more restricted areas, such as Kazakhstan and Mongolia, the creation of such a scheme presents difficulties. In contrast to the Pleistocene and Neogene, for which more or less workable schemes have been elaborated (Zazhigin, 1980), the Paleogene part of the Asian stratigraphic sequence as a whole, especially in Kazakhstan and Mongolia, remains poorly studied. Here, one meets difficulties not only in correlating isolated rodent remains, whose stratigraphic position is unclear in the so-called localities of the Dzhilanchik or Myneskesuek, with slightly more numerous Paleogene faunas of Mongolia, but also in correlating Kazakhstan faunas of similar ages (Chelkar-Teniz and Kyzyl-Kak) or the long-known but poorly studied faunas of Mongolia (Shand-Gol and Ergelyeen-Dzo).

In this respect, faunas of the Zaisan depression acquire a special importance. In this region, they characterize the time span from the early Eocene to the Pliocene (finds are known for nearly all formations), and they include more than simply rodents, languorous, and insectivores. Some of their stratigraphic levels yield up to 12 groups of vertebrates, invertebrates, and plants constituting representative complexes.

The Oligocene deposits in the Zaisan depression are represented by five stratigraphic units. These are the middle and upper parts of the Aksyir Formation, along with the Kusto, Buran, Oshagandy, and Nura formations. The Buran Formation is the focus of the following discussion. So far, the age of the formation has been based not only on its stratigraphic position in the sequence, but also on data of freshwater mollusks, remains of macroplants, and finds of large mammal *Cadurcodon zaissanensis* (Belijaeva, 1962; Borisov, 1963, 1983).

During the Saikan Expedition from 1961 to 1991 (Paleontological Institute, Russian Academy of Sciences, and the scientific-technical firm *Biokhron*), I collected for the first time new materials comprising various groups of animals and plants, which occur in association with each other and represent paleontological assemblages.

Materials were collected from a number of localities in the northern Munrak Range (basins of the Taizhuzgen and Mailibai rivers and the Dzhamangora–Chaibulak interfluve area). The assemblages include gyrogonites of stonewort algae, molluscs, ostracods, and vertebrates (fishes, amphibians, smaller reptiles, and turtles). Mammals are represented by bats, insectivores, rodents, lagomorphs, carnivores, and perissodactyls.

Localities situated between the Dzhamangora and Chaibulak rivers are confined to the lower part of the Buran Formation. Their position apparently corresponds to Bed 154 of the Kalmakpai River reference section, which contains fine-grained ferruginous sandstones (Borisov, 1963, p. 54). Finds in the Mailibai River basin come from a bed of compact clayey sands, which are similar to rocks of Bed 155 (Borisov, 1963, p. 55). Localities in the Taizhuzgen River basin are situated in basal beds of grey and green clays and have neither lithologic nor taphonomic analogues in the Zaisan depression. In other parts of the Zaisan depression, there are exposures of deposits of the Buran Formation that do not contain vertebrate remains. In these places, rich assemblages of macroflora, palynoflora, stonewort algae, and molluscs have been collected.

Finds of small mammals, rodents in particular, make it possible to correlate the known Oligocene faunas of Kazakhstan and then to compare them with faunas of Europe and Central Asia. The stratigraphic scheme of Borisov (1963, 1983) used here shows that the Bourn Formation corresponds to the second half of the lower Oligocene. Below, there is a list of the rodents distinguished in the Bourn Formation (Lychev and Shevyreva, 1994; Shevyreva, 1994).

Family Castoridae Gray, 1821
Agnotocastor devius sp. nov.
Steneofiber kazakhstanicus (Borissoglebskaja)
S. zaissanensis sp. nov.
S. shevyrevae sp. nov.
S. aff. *shevyrevae* sp. nov.
Family Cylindrodontidae Miller et Gidley, 1918
Cyclomylus mashkovae sp. nov.
Beatomys gloriadei sp. nov.
Family Cricetidae Rochebrüne, 1883
Cricetops collator sp. nov.
Leidymys asybaevi sp. nov.
Superfamily Ctenodactyloidea Zittel, 1893
Family Tamquammyidae Shevyrevae, 1983
Woodomys dimetron sp. nov.
Family Ctenodactylidae Zittel, 1893
Muratkhanomys velivolus gen. et sp. nov.
M. kulgayniae sp. nov.
M. djanarae sp. nov.
Roborovskia collega gen. et sp. nov.
Tataromys raeda sp. nov.
T. boreas sp. nov.

Before discussing the stratigraphic position of the Buran fauna and the age of the host deposits, some particular features of rodents and other groups are worth consideration along with additional evidence inferred from independent data on accompanying insectivores, lagomorphs, bats, and other vertebrates, invertebrates, and plants.

Comparison of rodents from the Buran assemblage with faunas of Kazakhstan, Central Asia, and Europe suggests that they are very similar at the generic level as far as the Chelkar-Teniz (Kazakhstan), Shand-Gol (=Tatal-Gol) (Mongolia), Taben-Bulak, and Saint-Jaques (China) localities are concerned. It should be emphasized that in comparison with Central Asian faunas of the Shand-Gol and Red Clays formations, where beavers are completely absent, the Buran assemblage is very rich in Castoridae. This is a common feature for the Buran assemblage and faunas of Kazakhstan and western Europe. The absence of beavers in Central Asia seems reasonable because of the evidently arid climate in this zone. Thus, according to its composition, the Buran fauna is intermediate between the aforementioned faunas of Kazakhstan and Central Asia. The similarity with the former is particularly evident in the identified Ctenodactylids: *Roborovskia collega, Tataromys raeda,* and *T. boreas*. Specimens of the genus *Tataromys* and similar forms have still not been found in the rodent faunas of Kazakhstan. The presented data leads us to the conclusion that the Buran Formation can be considered as a stratigraphic analogue of the Chilikty (= Chelkar-Nura or Indricotherium) Formation of Kazakhstan

or the Shand-Gol and Red Clays formations of Central Asia.

However, the morphological analysis of rodent dentitions from the above localities suggests somewhat different inferences. First, there are new genera and species in the Buran fauna. Second, despite the fact that some species are very similar in the structure of their cheek teeth to forms from the Chilikty Formation, it is quite clear that they are also similar to rodents from the Shand-Gol and Red Clays formations of China. Due to this similarity, *Beatomys gloriadei, Cricetops collator, Woodomys dimetron, Tataromys boreas, T. raeda*, and *Roborovskia collega* may be considered as taxonomic analogues of *Beatomys bisus* Shevyreva, *C. aeneus* Shevyreva or *C. dormitor* Matthew et Granger, *W. chelkaris* Shevyreva, *T. sigmodon* Matthew et Granger, *T. grangeri* Bohlin, and *T. deflexus* Teilhard de Chardin, respectively.

The abundance and diversity of beavers in the Buran fauna distinguish it from other known Eurasian faunas. This indicates favorable conditions for beavers and a variety of suitable biotopes, including those distant from large water basins. For instance, Wilson (1975) reasoned that *Agnotocastor* Stirton might be a fossorial form that inhabited river banks. It is also easy to assume that at least two stages of rodent fauna evolution occurred during the relevant period of the middle Oligocene.

The beaver *Agnotocastor* preserved P^3 in its dentition. This very primitive feature was inherited from ancestral forms. Thus, the presence of *Agnotocastor* enhances the archaic appearance of the Buran fauna. The distribution of this genus is restricted to the lower Oligocene of North America.

Due to the discovery of new species of Castoridae in the Buran fauna, the earlier viewpoint of Lychev (1974) on affinities, the existence time of beavers *Steneofiber* Geoffroi, and other aspects of the history of this family should be revised.

Beavers assigned to the genus *Steneofiber* have been reported many times from the Miocene deposits of Georgia and Kazakhstan (Burchak-Abramovich and Gabashvili, 1980; Lychev, 1982). *Steneofiber* sp., reported by Vinogradov (1936), should be excluded from this group, because this form is now considered identical to *Trogontherium minus* Newton (Lychev, 1974).

In Europe, beavers of the genus *Steneofiber* are characteristic of Oligocene deposits, though they are not very diverse: *S. butselensis* Misonne, *S. dehmi* Freudenberg, and *S. eseri* Mayor. Recently, *Steneofiber* sp. was discovered in an upper level of the lower Oligocene (Ünay-Bayraktar, 1987). European and Asian species of *Steneofiber* were most similar in the lower Oligocene, especially in its basal strata. However, among all species of this genus, only *Steneofiber kazakhstanicus*, which is known nearly everywhere in the Oligocene mammal faunas of Kazakhstan, is most similar to *S. butselensis* from the Hoogbutsel locality in Belgium. Its stratigraphic level and age are considered to correspond to the lower Oligocene (Rupelian) (Steininger *et al.*, 1982; Thaler, 1966).

It should be noted that, even though their teeth proportions are similar, the enamel of folds and fossetes is still more crenulated in *S. kazakhstanicus* than in *S. butselensis*. It seems that such features imply a more advanced evolutionary stage and a somewhat younger geological age of the Buran Formation in comparison with the Hoogbutsel fauna. In this case, it is reasonable to compare *S. kazakhstanicus* with *S. dehmi* from the Mine-de-Rochette fauna of Switzerland. *S. kazakhstanicus* appears to be a less advanced form than *S. dehmi* (Engesser *et al.*, 1984), because the crowns of cheek teeth are somewhat lower and more narrow. Therefore, the Buran fauna of the Zaisan depression is evidently older than the fauna of the Mine-de-Rochette. The age of the latter roughly corresponds to the level of the Bohningen zone or the Antoingt zone, that is, to the second half of the lower Chattian or the upper level of the middle Oligocene (Engesser *et al.*, 1984). Hence, it follows that the stratigraphic level of the Buran Formation is lower than the level of the Bohningen and Antoingt zones, and is higher than the level of the Hoogbutsel and Ronzon zones.

Species of beavers from the Buran Formation provided clear evidence regarding the faunal connections of Kazakhstan with Europe and North America. These connections were realized by intercontinental exchanges, which involved the Paleogene fauna of the Zaisan depression. Although abundant species of the genus *Steneofiber* can be regarded as autochthonous, the genus *Agnotocastor*, which never penetrated into Europe, is the real Oligocene migrant from faunas of North America. Judging by the remains of *Steneofiber*, beavers of the middle Oligocene of Turkey are more similar to European forms than to those from Zaisan localities. Consequently, during the Oligocene, Asia Minor was zoogeographically closer to the Mediterranean than any other province.

The species *Cyclomylus mashkovae* from the Zaisan fauna differs considerably in morphology from known species of *Cyclomylus* Matthew et Granger, such as *C. lohensis* Matthew et Granger and *C. turgaicus* Vinogradov et Gambarjan. They may even be regarded as representing an independent genus. It should be noted that *C. mashkovae* is less advanced evolutionarily than both *C. lohensis* and *C. turgaicus*. Apparently, the latter species should be assigned to a separate genus. Ancestral characteristics of *C. mashkovae* are expressed by the fact that the necks of the teeth narrow more distinctly. These teeth are funnellike but not cylindriform, as is typical of *C. lohensis* and *C. turgaicus*.

Cricetidae are poorly represented in the Buran fauna, but their remains also indicate connections with the western hemisphere. In its morphology of upper M^3, a new species *Leidymys asybaevi* shows clear similarity with North American forms of the genus *Leidy-*

mys Wood from the upper levels of the lower Oligocene of Colorado (Martin, 1980; Wood, 1936).

The genus *Muratkhanomys* occupies a special place among Central Asian Ctenodactylidae. The lower molars of all previously known genera, such as *Tataromys, Karakoromys* Matthew et Granger, *Leptotataromys* Bohlin, and *Dianomys* Wang (the lower molars of *Yindyrtemys* Bohlin are unknown), always have a variably developed oblique or curved mesolophid. It can be either barely visible, as in *K. decessus* Matthew et Granger, or massive and long, as in *D. quijingensis* Wang. When present, it has depressions that are variably open or closed and occur in the trigonid, behind the metaconid, or in front of the anterior pair of cusps, as in *T.* cf. *grangeri* (Kowalski, 1974, pl. 48, Fig. 5) and *Karakoromys* sp. (= *Leptotataromys gracilidens* Bohlin) (Bohlin, 1937, Fig. 101). On the contrary, the lower molars of *Muratkhanomys* have a different structure, though P_4 displays the typical ctenodactylid morphology. All the forms mentioned can be divided into two groups based on the structure of the molars. It is likely that the two groups correspond to two major trends in the evolution of the dentition of Ctenodactylidae, whose taxa have supergeneric status. It is possible that the Zaisan depression was isolated from the Central Asian subregion for a long time prior to the Oligocene. The evolution of Ctenodactylidae during the second half of the Paleogene was fundamentally different in Mongolia and China from that in the Zaisan depression. This is evident from numerous variations in the dental type of species, which was a principal feature of the Central Asian faunas that was controlled by the geography and climate of that time. At the beginning of the Oligocene, Asia was subjected to that more intense aridization than other regions in the eastern and western hemispheres. Of course, this would have an influence on the morphology of Asian animals, for example, some groups of rodents such as Cylindrodontidae (Shevyreva, 1971; 1974a, b; 1976a, b).

The Eocene–Oligocene boundary marks a new principle trend in the evolution of Ctenodactylidae that helped them to survive the arid climate of Central Asia. There was a set of rather diverse biotopes, which is evidenced by a number of adaptations that modified the general dental structure of the Oligocene Ctenodactylidae of Kazakhstan, Mongolia, and China.

Ctenodactyloid rodents of the Buran fauna differ from the Eocene forms of the Chakpaktas, Obaily, Sargamys, and Konur-Kura formations in that their degree of morphological advancement is higher despite the preservation of the general type of dentition structure or, more precisely, the occlusial surface. Whereas many ctenodactyloids (*Tamquamys* Shevyreva, *Petrokozlovia* Shevyreva, *Geitonomys* Shevyreva, *Bolosomys* Shevyreva, *Borissomys* gen. nov., *Ekaterinomys* gen. nov., *Juniperusomys* gen. nov., and others) have bunodont teeth, the occlusial surface of the Buran forms *Muratkhanomys, Tataromys boreas, T. raeda,* and *Roborovskia collega* has a lophodont structure. These changes in the dental system imply saltation, i.e., the radiation and formation of higher supergeneric taxa. This process reflects one of the main natural laws: the law of parallel development. Among all herbivorous animals, rodents provide the best example of parallelism. In this case, one of the manifested stages of morphological evolution is the formation of a high crowned tooth without roots (hypsodont) but with a pectinate flattened chewing surface, which is well-adapted for grinding cellulose food (Shevyreva, 1976b). Dynamic changes in the composition of Zaisan Ctenodactylidae from their Eocene to Oligocene (Buran) rodent faunules distinctly suggest this.

Ctenodactylidae from deposits of the Buran Formation show that the Zaisan faunas had temporal and territorial connections with those of Kazakhstan, Mongolia, and China. However, in the Buran Formation, the group is represented by somewhat less abundant remains than in the faunas of Mongolia and China, where Ctenodactylidae represent one of the main components along with lagomorphs and rodents of the family Cylindrodontidae. On the basis of faunal connections, the Buran fauna may be regarded as an approximate analogue of the faunas from the Shand-Gol and Red Clays formations, corresponding to the second half of the lower Oligocene. Moreover, the genus *Muratkhanomys* gives the Buran fauna a somewhat more archaic appearance, because species of the genus display both advanced characteristics, such as lophodont teeth, and primitive features, such as relatively low crowns, weak and more differentiated roots, and low longitudinal crests in P_4. These features suggest the affinity of the group with the Eocene Ctenodactylidae.

A comparison of the Buran and Oligocene rodent faunas of Kazakhstan proves that the former is from a lower stratigraphic level than the fauna from the Chilikty (= Chelkar-Nura, Indricotherium) Formation of the middle Oligocene. This conclusion is supported by the presence of *Agnotocastor devius, C. mashkovae, Muratkhanomys* spp., and *Roborovskia collega* among the Buran forms that are more archaic than the Chelkar-Teniz or Kyzyl-Kak faunas. A further comparison also shows a time difference between the latter and the Shand-Gol, Tatal-Gol, Taben-Bulak, Saint-Jaques, and other faunas of Central Asia (Shevyreva, 1976a). The Oligocene faunas of Kazakhstan include some elements that are more archaic than those in the Central Asian faunas. Hence, the data on rodents indicate that the Chilikty and Shand-Gol formations are directly correlative with the Buran Formation of the Zaisan depression only within a restricted time interval, namely, in the middle part of Oligocene.

As the Oligocene rodents of Kazakhstan are poorly studied in all known mammal faunas, it is impossible to analyze them in detail and to understand their time relationships. It is not inconceivable that these faunas are different in age. Only the following middle Oligocene sequence of faunas (from older to younger) seems to be warranted: the Buran fauna → faunas of the Turgai

Depression in Kazakhstan ⟶ faunas of the Shand-Gol and Red Clays formations. It should be noted that the Buran fauna occupies the lowest stratigraphic level.

A similar approach in the analysis of other fossils from the Buran complex that have been independently studied by other scientists provided very interesting results. These results are concordant with the data on rodents and suggest, in addition, that the stratigraphic position of the Buran fauna should be lowered to the top of the lower Oligocene.

Taxa of the order Chiroptera (cf. *Myotis* sp., family Vespertilionidae) were first found in the Paleogene deposits of Kazakhstan (Gabunia & Gabunia, 1987b). Represented by solitary forms, they are, thus far, useless for a reliable substantiation of the stratigraphic position of the fauna from the Buran Formation. Their similarity with species of the Oligocene genus *Myotis* from western Europe provided only a weak basis for reliable systematics, especially because the present state of our knowledge on the Oligocene myotis species is insufficient. This hampers a correlation between the Buran and western European faunas.

The Insectivora of the Buran fauna are represented by aundant and diverse forms (Gureev, 1979; Gabunia & Gabunia, 1987a; Gabunia *et al.*, 1983):

Family Tupaiidae Gray, 1825

Tupaiidae gen. et sp.?

Family Erinaceidae Fischer von Waldheim, 1817

Subfamily Galericinae Pomel, 1848

Galerix sp.

Pseudoneurogymnurus shevyrevae Gureev

P. zchickvadzei Gureev

Galericinae gen.

Subfamily Erinaceinae Fischer von Waldheim, 1817

Amphechinus cf. *rectus* Matthew et Granger

Amphechinus sp.

Tupaiodon sp. cf. *morrisi* Matthew et Granger

Ictopidium cf. *tatalgolensis*

Ictopidium sp.

Family Talpidae Fischer von Waldheim, 1817

Subfamily Uropsilinae Thomas, 1912

Uropsilinae gen. et sp. cf. *Uropsilus cipes* M.-Edwards

Subfamily Desmaninae Thomas, 1912

Desmaninae gen.

Desmanella sp.

Subfamily Talpinae Thomas, 1912

Urotrichini cf. *Urotrichus talpoides* Temmincki

Talpinae gen. et sp.? cf. *Parascaptor leucura* Blith

Talpidae inc. sedis

Family Soricidae Fischer von Waldheim, 1817

Subfamily Soricinae Fischer von Waldheim, 1817

Gobiosorex cf. *kingae* Sulimski

Heterosoricinae gen.

The Buran fauna is rich in specimens of hedgehogs and shows the first occurrence of true moles, together with desmans, and the further development of shrews.

Because *Amphechinus* cf. *rectus*, *Tupaiodon* cf. *morrisi*, *Gobiosorex* cf. *kingae*, and *Ictopidium* cf. *tatalgolensis* occur in the Buran fauna, it resembles the fauna of the Shand-Gol Formation (Mongolia); thus, both seem to be similar in age, i.e., they originated in the middle Oligocene.

In comparison with rodents, insectivores are more precise indicators of global events that influenced all mammals, for instance, changes in their distribution during the Cenozoic. In the Oligocene, Galerecinae were common faunal elements in the territory of the modern Palearctic. In the Miocene, they migrated to the Ethiopian region, and in the Anthropogene, they penetrated into the Indo-Malayan province. For example, hedgehogs of the genus *Amphechinus* and hairy hedgehogs of the genus *Galerix*, which are known in the Buran fauna, are recorded in the Miocene of Africa and even North America. Other forms, such as *Pseudoneurogymnurus* or the oldest hedgehogs of the Tupaiidae family, became extinct in the Zaisan depression; however, their closest relatives *Echinosuchus* Blainville, *Hylomys* Müller et Schlegel, and others along with diverse tree-shrews *Anatana* Lyon, *Dendrogale* Gray, and others inhabited Indostan and islands of the Malay Archipelago. Insectivores of the Buran fauna are represented by a large group of families indicative of a wide range of specialization. Moreover, the morphological diversity within each family, evident from the list of forms, reflects variable ecological adaptations.

The presence of Tupaiidae gen. et sp.? is noteworthy. These tree-shrews live in arboreal and terraneous habitats of tropical forests, rain forests, and brushwoods.

Gymnures or hairy hedgehogs (Galerecinae) inhabit not only wooded landscapes, but cultivated ones as well; the presence of water is obligatory. Many forms are able to feed in the water and have made adaptations for fossoriality, swimming, and cursorial locomotion.

True hedgehogs (Erinaceidae) inhabit all landscape types and generally are not very agile animals.

Shrews (Soricidae) are adapted more than other insectivores to fossoriality, swimming, swift cursorial movements, and bouncing locomotion. Their dentition is considered to be the most specialized and is adapted to various diets. Shrews always live in humid biotopes.

Moles (Talpidae) are chiefly adapted to fossorial and aquatic life. Diverse modifications in their trends of development are represented by a variety of biological types that formed during the Tertiary.

In general, the insectivores from the Buran fauna were dwellers of humid tropical forests and bushes along the banks of water basins: tree-shrews, gymnures (Galerecinae), water moles (Desmaninae), Japanese

shrew-moles (Urotrichini), Asiatic shrew-moles (Uropsilinae), and shrews.

Lagomorphs (Lagomorpha) are rather abundant in the Buran fauna (Erbaeva, 1994).

Family Leporidae, Gray, 1821
Subfamily Mytonolaginae Burke, 1941
Gobiolagus cf. *major* Burke
Gobiolagus sp.
Subfamily Agispelaginae Gureev, 1960
Procaprolagus sp.
Agispelagus cf. *simplex* Argyropulo
Family Palaeolaginae Dice, 1929
Subfamily Palaeolaginae Dice, 1929
Desmatolagus cf. *gobiensis* Matthew et Granger
D. sp. (aff. *gobiensis*)
D. aff. *robustus* Matthew et Granger
Desmatolagus sp. A

The species composition of lagomorphs in the Buran fauna is identical to the lagomorph assemblage from the Shand-Gol Formation, and some Zaisan species are very similar to Mongolian forms. However, some taxa of the Buran fauna, for instance, species *Gobiolagus* cf. *major* and *Gobiolagus* sp., are unknown in the Shand-Gol fauna. Species *Gobiolagus* and *Procaprolagus* coexisted in the Buran Formation of the Zaisan Depression, and this lagomorh assemblage is similar to that from the Shara-Murun locality (China, Inner Mongolia). The latter fauna has a late Eocene age (Burke, 1941). From this viewpoint, the lagomoph fauna of the Buran Formation is apparently somewhat older than the Shand-Gol fauna.

Ecologically, the Buran lagomorph fauna differs from the fauna of the Shand-Gol Formation. The Buran fauna completely lacks true Ochotonidae. Presently, pikas represent the biological type adapted to mountainous and rocky areas or to plains and piedmonts with rather severe conditions.

The complete absence of pikas, which apparently had a colonial mode of life like the modern species, was probably not caused by a taphonomic effect, but was the result of the absence of their favorable habitats at that time.

The distribution area of the Paleogene pikas was presumably outside of the Zaisan depression.

Large mammals in the Buran fauna are represented by the remains of carnivores and ungulates (Gabunia *et al.*, 1983): cf. *Cynodictis,* Hyaenodontidae gen. indet., Perissodactyla (*Cadurcodon zaissanensis* Bel.), and Artyodactyla (*Miomeryx* sp.).

Representatives of the genus *Cynodictis* are known from the upper Eocene–lower Oligocene deposits of Europe and from the lower-middle Oligocene sediments of Asia (Flerov *et al.*, 1974).

Genera *Cadurcodon* Kretzoi and *Miomeryx* Matthew et Granger are typical of fauna from the Ergelyeen-Dzo Formation (= Ergil'-Obo) of Mongolia. The age of this fauna corresponds to the early Oligocene (Belijaeva, 1962; Gromova, 1960; Devyatkin, 1981).

Regarding the stratigraphic distribution of chevrotains of the genus *Miomeryx*, it should be noted that tragulid remains from the upper Oligocene locality Benara (Belijaeva and Dubrovo, 1962) that were tentatively assigned to this genus actually represent species of a different taxon (Gabunia, 1964).

Hence, the generic composition of large mammals from the Buran Formation is correlative with that of the Ergelyeen-Dzo fauna. Therefore, according to the fauna of large mammals, the age of the Buran Formation should not be younger than the early Oligocene.

Both the swamp rhinoceros *Cadurcodon zaissanensis* Belijaeva and the tragulid *Miomeryx* inhabited water-rich forests with humid and swamp conditions (Belijaeva and Dubrovo, 1962).

The identitical generic composition of carnivores and ungulates suggests natural territorial integrity of the Zaisan Depression and Mongolia, which were identified as a zoogeographic unit that existed in the first half of the Oligocene in the North Asian subregion (Flerov *et al.*, 1974). Finds of *Cynodictis* in Europe indicate a certain faunal exchange between the European and Asian provinces during the Eocene–Oligocene boundary period.

Another group of vertebrates in the Buran fauna is represented by fishes *Amia* sp., *Eodiptychus* sp., *Esox* sp., *E. aralensis* Sytchevskaya, *Catostomidae* gen. indet., *Rutilus* sp., *Tribolodon* sp., *Zaissanotinca cristidens* Sytchevskaya, *Parabarbus* sp., *Schizothorax* sp., and Cyprinidae gen. indet., which were identified by Sychevskaya (1986).

Modern data on this faunal group are insufficient for correlating the Buran ichthyofauna with any fish assemblage known in the territory of the former USSR or in the adjacent regions of Mongolia and China. The freshwater Buran ichthyofauna is represented by primitive benthophages of the Catostomidae family (Sychevskaya, 1986). These forms live in well-aerated water, which supports the possibility of a transgression during the Buran time. This is especially indicated by the gradual replacement of shallow water species by Cyprinidae (*Rutilus* sp., *Tribolodon*, and others), which became abundant in upper beds.

Amphibians (Anura) of the Buran fauna were determined as Pelobatidae gen. indet. (Gabunia *et al.*, 1983).

The more abundant Buran reptiles are represented by *Ergilemys insolitus saykanensis* Ckhikvadze, *Crisemys index* Ckhikvadze, *Chelidropsis minax* Ckhikvadze, *Plastomenus* sp., *Tryonyx* sp., *Planiplastron* sp., and *Ocadia* sp. (Chkhikvadze, 1972, 1973; Chkhikvadze and Shevyreva, 1975).

Turtle fauna of the Buran Formation includes chiefly endemic species. However, *Ergilemys insolitus* was also found in the Ergelyeen-Dzo fauna (Mongolia); this shows that at least the basal beds of the Buran

Formation correspond to the Ergelyeen-Dzo Formation of the latest Oligocene.

The most abundant are the freshwater turtle species, associated with a single land form *Ergilemys insolitus*. Such proportions indicate the prevalence of humid biotopes in the landscape bordering the lake situated in the south of the Zaisan depression.

All aquatic turtles of the Buran fauna differ greatly in their biological features. Judging from modern *Chrysemys, Ocadia,* and *Plastomenus*, forms like *Chrysemys index* should inhabit slow-flowing or stagnant waters full of vegetation, which was their main source of food. However, forms like *Planiplastron*, though not very active swimmers, still prefer swift mountainous rivers. Similar to their present-day representatives, *Planiplastron* moved along rocky bottoms by effectively clutching with their claws at wrinkles in the relief, and they hunted fishes and aquatic invertebrates. Soft-shelled and loggerhead turtles of the genus *Chelydropsis* are relatively large and active predators. Another form, *Plastomenus*, presumably occupied the same ecological niche as modern North American forms of Tryonichidae, feeding on aquatic insects and crustaceans that are able to burrow into mud.

The morphological similarity in the plastron structure of *Plastomenus* and *Tryonyx* species reflects ecological parallelism. Therefore, *Plastomenus* adapted to aquatic habitats to a greater extent than other turtles of the Buran fauna.

Diverse and abundant water turtles of the Buran fauna are indicative of favorable aquatic biotopes. At the same time, dwarf forms are recorded among the *Plastomenus* species. This could be a response to the deterioration in living conditions that was initiated at that time.

The endemism of the Buran turtles seems to have resulted from the relative isolation of the Zaisan depression within the Paleoarctic provinces despite the evident land connections with regions of Central Asia, such as Mongolia.

The distinguished invertebrates (Mollusca) are *Sphaerium proscaldianum* Mart. et Tolst., *Amesoda martinsoni* Tolst., *A. kustovica* Tolst., *Musculium proelacustris* Tolst., *Pisidium jugatum* Tolst., *Bithynia buranensis* Tolst., and *Bithynia* sp. (Martinson and Tolstikova, 1969; Tolstikova, 1973, 1980). The molluscan fauna of the Buran and other Zaisan formations have no analogues in other regions and it is useless to constrain the age of the Buran Formation. The fauna lacks terrestrial forms, which suggests that only aquatic biotopes existed in the Buran time. Buran mollusks are either shallow-water or swamp-living eurybiontic bivalves, such as *Sphaerium, Amesoda,* and others, or gastropods that endured periodic drainage in littoral zones.

Ostracods also have no significance in discussions about the stratigraphic position of the Buran fauna. However they are rather diverse: *Iliocypris errabundis* Mand., *I. manasensis, I. manasensis* var. *cornae* Mand., *Kasinina beliaevskyi* Mand., *Candoniella albicans* (Brady), and *Eucypris grandis* Bodina (Stankevich, 1980).

Stonewort algae are represented by *Hornichara kazakstanica* Maslov, *H. ashutasica* Romash., *Grambastichara vasilenkii* Romash., *G. tornata*, (Reid et Groves), *Raskyechara claerae* Romash., *R. peckii* (Rasky), *Amblyochara orbae* Romash., and *Harrisichara maslovi* Romash. (Kyansepp-Romashkina, 1980).

Owing to the presence of *Hornichara kazakstanica*, the algae assemblage is correlative with that of the Chelkar-Nura (= Chilikty) Formation in the Chelkar-Teniz locality of central Kazakhstan. Most students consider the latter formation to be an analogue of the Shand-Gol Formation, and, thus, corresponding in age to the middle early Oligocene. However, an alternative interpretation (Shevyreva, 1976a) favored the second half of the early Oligocene (the first half of the middle Oligocene of former scales).

According to the stonewort algae, the age of the Buran Formation corresponds to the entire Oligocene interval (Kyansepp-Romashkina, 1980).

In the Buran time, stonewort algae were widespread and abundant along the southern and southwestern shorelines of the lake. This was possible because the conditions of the muddy bottom were undisturbed, which is typical of many zones in the depression. The lack of ornamentation in gyrogonites may be an indication of insignificant water dynamics (Kyansepp-Romashkina, 1980).

The composition of higher plants is known both from the remains of macroflora (Akhmet'ev, 1985; Il'inskaya, 1962, 1986; Il'inskaya *et al.*, 1983) and from palynological data (Boitsova and Panova, 1973). The list of the former is very extensive: *Tsuga* sp., *Magnolia* cf. *ingliefildi* Heer, *Sassafras ferretianum* Massal, *Menispermum* sp., *Cercidiphyllum crenatum* (Ung.) Browm, *Fotergilla turgaica* Grub., *Liquidambar europea* A.Br., *Platanus* sp., *Eucommia palaeoulmoides* Baik., *Eucommia* sp., *Ulmus drepanodonta* Grub., *U. carpinoides* Goepp., *Zelkova zelkovifolia* (Ung.), *Celtis irtyshensis* Akhm., *Alnus* sp., *Carpinus* ex gr. *cordata* Blume, *Carpinus* sp. 1, *Carpinus* sp. 2, *Ostrya antiqua* Grub., *Corylus jarmolenkoi* Grub., *Myrica* sp., *Carya cordioides* Il., *C. borissovi* Il., *Carya* sp. 1, *Carya* sp. 2, *Cyclocarya cycloptera* (Schlecht.) Il., *Populus* aff. *jarmolenkoi* Il., *P. arnaudi* Il., *P.* ex gr. *balsamoides* Goepp., *Populus* sp., *Salix* sp., *Tilia irtyshensis* (Schap.) Baik., *Maakia zaissanica* Akhm., *Leguminosites* sp., *Cotinus genuina* Il., *Acer monoides* Shap., *A. tricuspidatum* Ett., *A. neuburgae* Baik., *Acer* sp. 1, *Acer* sp. 2, *Alangium aequalifolium* (Goepp.) Krysht., *Celastrus* sp., *Ampelopsis* sp., *Ungnadia kryshtofovichi* Il., *Vitis* sp. 1, and *Vitis* sp. 2.

The palynological spectrum of the Buran Formation is similar to the *Carya spackmania–Tilia grandireticulata* assemlage known from the Ashcheairyk Formation of the Turgai depression. The latter palynospectrum is the early-middle Oligocene in age (Il'inskaya *et al.*, 1983).

Hence, the plant remains show that the age of the Buran Formation may correspond to the early Oligocene.

Based on A and B assemblages of deciduous flora (Akhmet'ev, 1985; Il'inskaya, 1986), two groups can be recognized. The first consists of valley plants (microphyllous poplars, Juglandaceae, and Fabacea), and the second consists of slope vegetation, including hydrophilic and thermophilic forms (*Liquidambar, Magnolia*, and others). The flora of the Buran Formation was transitional between plant assemblages of subtropical and warm-moderate zones. It marked the terminal period in the expansion stage of the most thermophilic plants of the Turgai type (Il'inskaya, 1962).

In conclusion, it should be noted that the fauna of the Buran Formation, together with plant remains, is hardly identical to fossils known from the Shand-Gol or the Chilikty formations. Many archaic mammals occurring among the rodents, insectivores, and ungulates suggest that the Buran fauna is more similar to the somewhat older early Oligocene faunas of Europe and Central Asia. In general, the stratigraphic position of the Buran fauna is considered here to better fit the upper part of the lower Oligocene, i.e., it corresponds to a level that is higher than the Hoogbutsel and Ronzon zones but lower than the Bohningen and Antoingt zones (Berger *et al.*, 1987; Ünay-Bayraktar, 1987). Thus, the age of the Buran Formation is either the earliest middle Oligocene, according to former scales, or it corresponds to the begining of the second half of the early Oligocene, according to the modern scale.

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Zoogeography of the Early Oligocene Basins of West Eurasia Based on Bivalve Mollusks

S. V. Popov

Paleontological Institute, Russian Academy of Sciences, Profsoyuznaya ul. 123, Moscow, 117647 Russia

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Abstract – As in the Eocene, the boundary between the Tethyan and North Europe during the early Oligocene remained the main zoogeographical boundary of West Eurasia, although it was considerably displaced southward. The Tethys was still a tropical region, but its fauna, which was mainly inherited from the Eocene, became impoverished. The North European molluscan fauna was considerably renewed and acquired a boreal-subtropical aspect. The German and East Paratethyan provinces were distinguished within this realm. A uniform endemic brackish-water fauna appeared in the Paratethys after its closing during the late Rupelian.
Key words: *Biogeography, Paleogene, mollusks, Tethys, Paratethys.*

In the late Priabonian–early Rupelian (s.l.), significant paleogeographic and climatic changes occurred that greatly affected the composition and zoogeographic distribution of the molluscan faunas. Changes in atmospheric and oceanic circulation caused subglobal climatic cooling. As a result, the Eocene "nonglacial" climate changed to a "glacial" one, and eustatic fluctuations intensified. The collision of the African and Indian plates with Eurasia caused the Tethys Ocean to close. Subglobal regression and Alpine orogenesis resulted in the emergence of the vast territories of West Siberia and Turgai. The Dinarides, Pontides, Lesser Caucasus, and Elbrus Range became a single continental massif, which separated the Paratethys from the Tethys. All these events led to considerable impoverishment and renewal of the molluscan composition, which were pronounced differently in the Tethyan and North European regions.

The methods and materials used in this article were discussed in a previous publication (Popov, 1994). Although the Rupelian molluscan fauna is less diverse than the Priabonian one, it is better represented in studied collections and was recently revised (Popov *et al.*, 1993). This allows wider use of the data on the species composition of assemblages in addition to those on the generic and subgeneric composition (Table 3).

ASSEMBLAGE CHARACTERISTICS

North Africa

Unlike those of the Priabonian faunas, records of the early Oligocene molluscan faunas of Egypt (Cuvillier, 1930) and Algeria (Flandrin, 1938) are very sparse. According to these publications, the fauna of the African coast had lost its originality by the Oligocene: all 13 genera reported from this area were widespread and also occurred in Europe, and almost half of the species are known to have existed in South Europe since the Eocene. Among them, there are characteristic North Tethyan forms, such as *Chlamys biarritzensis, Pecten arcuatus,* and *Pycnodonte brongniarti.*

South Europe

Representative early Oligocene fossil assemblages have been recorded in northern Italy (the Castelgomberto locality, Coletti *et al.*, 1973; the Vichenzo locality, Piccoli *et al.*, 1977). Bivalves were still represented in this area by very thermophilic tropical-subtropical assemblages with *Trisidos, Crassatella* s.s., *Cardium, Vepricardium,* and *Corculum* (Fig. 1). This fauna was mainly inherited from the fauna of the late Eocene: 37 of the 73 bivalve species found in Venice also appear in the Eocene (Piccoli *et al.*, 1977). However, the Priabonian assemblages also contain species that inhabited only the North European seas: *Barbatia nysti, Anadara sulcicosta,* and *Astarte kickxi.*

The Rupelian mollusks from southern Bulgaria (Karagyuleva, 1964) are also represented by thermophilic forms, including characteristic Tethyan species: *Chlamys subdiscors, Ch. biarritzensis, Pecten arcuatus, Spondylus cizalpinus, S. bifrons, Pycnodonte brongniarti, Pseudomiltha gigantica,* and *Macrosolen hollowaysii.* More than two-thirds of these species were inherited from the Eocene. However, this fauna is noticeably less diverse than the late Eocene one, and it also contains North European species (*Barbatia nysti, Pelecyora westendorpi, Codalucina tenuistria, Limatula eximia,* and *Pycnodonte quateleti*).

Transcaucasia

The early Oligocene molluscan fauna of Armenia (the Shorakbyur Formation, "Horizon with *Variarmussium fallax*") also includes many characteristic Tethyan genera and species. More than two-thirds of the forms

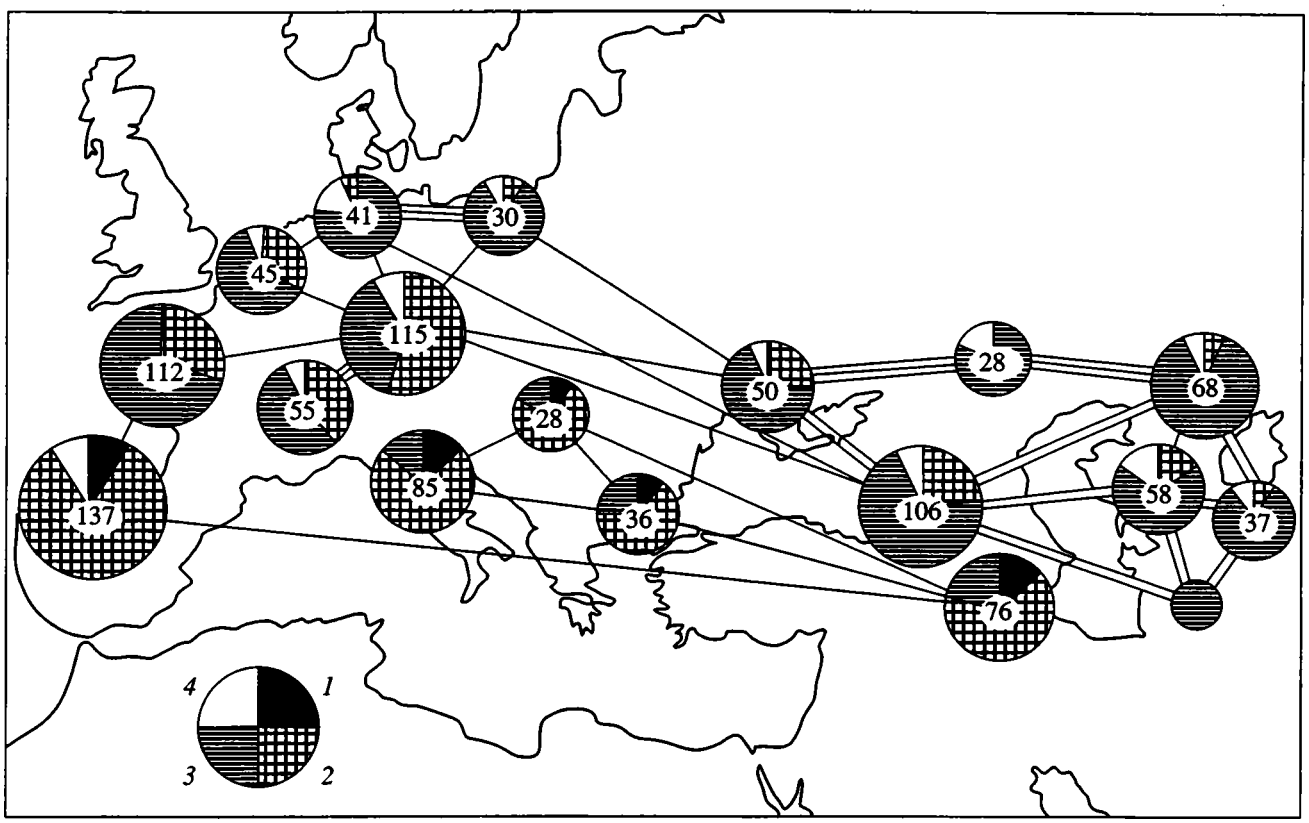

Fig. 1. Circle diagrams showing the diversity and thermophilic affinities of the Rupelian bivalve assemblages. The circle sizes are proportional to the species number in the center of the circle.

1 – species of tropical genera, *2* – species of tropical and subtropical genera, *3* – widespread species, *4* – predominantly boreal species. Lines connecting the circles indicate the degree of similarity of the assemblages.

found there were inherited from the Eocene. As in the above-mentioned faunas, new forms appeared along with North European species: *Nucula sulcifera, N. compta, Nuculana perovalis, N. crispata, Bathyarca bellula, Dimya fragilis, Pycnodonte quateleti, Megaxinus omaliusi,* and *Thyasira nysti.* The Transcaucasian molluscan fauna is the most similar in composition to the southern Bulgarian fauna (11 common species) and differs significantly from the very diverse Rupelian fauna of southern Georgia. The latter fauna contains more than 100 species of bivalves, only 8 of which are found in the Armenian assemblages. The great diversity of the Georgian fauna, as compared with the Armenian one, is undoubtedly secondary and can be explained by the fact that this fauna is better preserved and studied. However, the generic composition and the presence of Tethyan genera and species suggest that the fauna of the basin that existed in Armenia was notably more diverse and thermophilic.

There are some Tethyan species (*Modiolus rutimeyeri, Pecten arcuatus,* and *Chlamys meissonieri*) in the Georgian assemblages as well, but they are rare and inherited from the Eocene. Therefore, their occurrence does not provide evidence for the existence of a direct connection between Transcaucasia and the Tethys. This connection probably ended in the middle of the Priabonian. Almost half of the Georgian species are also found in the northern assemblages of the Black Sea region, Mangyshlak, and Ustyurt, although this fauna is characterized by the presence of more thermophilic mollusks.

After the freshening of the East Paratethyan basin, including its Transcaucasian part, during the second half of the early Oligocene, the marine fauna in Georgia was replaced by brackish-water assemblages with endemic genera and species.

Central Europe

In Romania, an early Rupelian shallow-water molluscan assemblage was found in the Hoia and Mera beds of the Transylvanian depression (Moisescu and Meszaros, 1974; Rusu, 1977). Although this assemblage is not diverse, it is indicative: it includes thermophilic bivalves belonging to peculiar Tethyan genera and species (*Chlamys biarritzensis, Spondylus cisalpinus, Macrosolen hollowaysi, Periglypta aglaurae, Callista villanovae,* and *Pelecyora verneuili*) as well as species of tropical genera, such as *Pinna* s.s. and *Discors.* The assemblage is mostly inherited from the Eocene: the share of Oligocene species (including

gastropods) amounts to only 14% (Moisescu and Meszaros, 1974). The majority of the species are in common with those of the Eocene and Oligocene assemblages of Italy and southern Bulgaria. These data are both reliable (confirmed by my examinations of sections and collections) and unique, because a thermophilic Tethyan molluscan assemblage of this type was not found in any other Rupelian deposits in Central Europe. Due to this fact, the Hungarian researcher Baldi (1986) concluded that a direct connection between the Central European basins and the Tethys did not exist at the very end of the Eocene and the beginning of the Oligocene. However, the composition of the Transylvanian fauna disproves this conclusion.

In Hungary, there are only relatively deep-water lower Rupelian deposits, which either lack mollusks or contain very impoverished assemblages that can not be used for comparison. The thermophilic Rupelian mollusks of Moravia (Oppenheim, 1922) and Switzerland (the Berner Jura locality, Kissling, 1896) did not include characteristic Tethyan species. The more diverse Swiss assemblage was similar to the Mainz basin fauna: 41 of the 55 bivalve species are in common with the latter. Both assemblages contained *Arca biangula, Ctena squamosa, Codalucina tenuistria, Spondylus tenuispina,* and other thermophilic forms (Fig. 1).

In Central Europe, numerous records of brackish-water upper Rupelian sediments appear with peculiar assemblages of nannoplankton, ostracods, and mollusks. These sediments were found in Hungary (the upper part of the Tard Clays), Romania (the Bizusa and Ileande beds of Transylvania), and Slovenia (Rusu, 1983; Baldi, 1986). The Transylvanian sediments yielded bivalves *Janschinella garetzkii, Ergenica* sp., *Urbnisia lata, Merklinicardium apostolovense,* ?*Korobkoviella* sp., *Cerastoderma serogosicum*, and *C. lipoldi* (my identifications of Rusu's specimens), and the Hungarian sediments contain *Janschinella* sp., *Urbnisia* sp., and *Cerastoderma lipoldi*. This allows these assemblages to be correlated unambiguously with the Solenovian (late Rupelian) ones of the East Paratethys.

West Europe

The well-studied Stampian mollusks of Aquitaine are very diverse (140 species, according to Cossmann, 1921; Saubade, 1970). The very thermophilic generic composition, including some tropical taxa, and the presence of specific Tethyan genera and species [*Lima* s.s., *Megaxinus* (*Megaxinus*), *Cardita* s.s., *Trachycardium, Discors*, and *Periglypta aglaurae*] suggest that a connection existed between the West European basins and the Tethys. However, the entire molluscan assemblage has a peculiar character, and few of its species are in common with those of the assemblages from South Europe and the Parisian basin. It is difficult to determine how many taxa were inherited from the Eocene, because the late Eocene fauna of Aquitaine is impoverished, and the Stampian mollusks are dominated by the forms that were first described here.

A faunal exchange, complicated by the climatic barrier, probably existed between Aquitaine and the Parisian basin through the Atlantic Ocean. This is evidenced by the presence of common species and the transitional character of the Bretagne fauna (the Rennes locality; Saubade, 1970).

The rich Stampian bivalve fauna of the Estampes Sands in the Parisian basin includes 112 species (Cossmann and Lambert, 1884; Furon and Soyer, 1947), but the thermophilic forms are less diverse. The share of Eocene species amounts to one-third, i.e., less than that in the South European assemblages. The fauna consists of predominant endemics and widespread species typical of the Rupelian (*Siliqua nysti, Angulus nysti, Callista splendida, Chlamys picta, Nemocardium tenuisulcatum, Spondylus tenuispina*, and others). There are 33 species in common with the Mainz basin fauna, and only 10 species are in common with the Belgian assemblages. This implies the existence of direct faunal links with the Bavarian basin via Alsace and, probably, the absence of such links with the North Sea basin (Saubade, 1970). Tethyan species and genera were absent in this fauna.

The Hamstead Beds of the Hampshire basin in England, which are correlated with the Stampian deposits of the Parisian basin, contain impoverished (24 species), mainly endemic fauna. This fauna includes freshwater forms and some marine species identical to those of the Parisian basin.

The Rupelian deposits of the Mainz basin yielded diverse bivalve fauna (115 species, which were recently revised thoroughly) (Sandberger, 1858 - 1863; Neuffer, 1973). Many thermophilic genera and species occurred (*Arca, Acar, Plicatula, Limaria, Ctena, Chama, Tivelina, Clavagella,* and others). This made the fauna comparable to the Stampian assemblages of the Parisian basin and the Rupelian ones of Switzerland and Transcaucasia and distinguishable from the less diverse northern faunas.

Nearly 60 species of bivalves (Glibert and Heinzelin, 1954) are known from the Rupelian of Belgium (the Berg Sands, clays with *Nucula comta*, and the Boom Clays). Thermophilic elements in this fauna are more diverse than in other molluscan assemblages from the lower Oligocene of the North Sea basin: *Glycymeris* (*Tucetilla*) *angusticostata, G.* (*Glycymeris*) *obovata, Codalucina tenuistria, Arcopagia, Isognomon, Spondylus,* and *Caestocorbula* were reported only from this area. These forms make the fauna of Belgium similar to that of the Mainz basin and Switzerland. This diversity can be attributed to both the specific shallower water facies of Belgium and the proximity of the area to the passage into the Central European basins through the Rhein graben, Mainz basin, and Alsace (Saubade, 1970, Fig. 3). Late Eocene species constituted less than half of the entire assemblage (Glibert and Heinzelin, 1954).

Very similar fauna known from the Rupelian clayey deposits of the Netherlands (Albrecht and Valk, 1943), Denmark (Harder, 1913), northwestern Germany (Haas, 1889; Hartkopf, 1979), northeastern Belgium (Janssen, 1981), and northern Poland (Wozny, 1965; Jakubowsky, 1983) is less diverse than the above-discussed assemblage. *Paleotaxodonta, Thyasira, Astarte,* and *Pectinidae* are predominant. But these faunas cannot be compared reliably, because their features were dependent on the sedimentation environments, which were inhabited by mollusks.

East Europe

Assemblages of the early Oligocene (Pshekhian) mollusks derived from wells and sections in the Black Sea depression, Volga–Don region, Crimea, and Ciscaucasia are generally very representative (74 species and 53 genera and subgenera), despite the low diversity of some of them, and have many taxa in common. Generally, they are soft-ground assemblages, which, like the North European ones, contain predominant *Paleotaxodonta* and diverse *Pectinidae, Carditidae, Astartidae,* and *Cuspidariidae* (Table 3). The share of subtropical elements does not exceed 15 - 20%. This fauna shows the most similarity with mollusks from the Transcaspian region [the Preston coefficient values (K) are 0.33 for genera and 0.54 for species] and Transcaucasia (0.46 and 0.66, respectively), despite the climatic and facies differences between these areas. Outside the East Paratethys, this fauna appears to be the most similar to the North Sea bivalve assemblage (Tables 1 and 2). The lack of records of late Priabonian shallow-water fauna of East Europe makes it impossible to determine how many taxa were inherited from the local Eocene assemblages. The species known from the Eocene deposits account for 50%; most of them are similar to the Latdorfian forms of the North Sea.

Mollusks from the upper Rupelian (Solenovian) deposits are represented by endemic brackish-water assemblages.

Transcaspian Region

Data on the early Oligocene mollusks of the Transcaspian region are complete and derived from the Ashcheairyk Formation of northern Ustyurt and sections and wells of southern Mangyshlak. Due to the facies diversity of the Ashcheairyk Formation, mollusks that inhabited soft grounds occur along with those that preferred sandy substrates. Although this fauna is diverse (68 species) and well-preserved, the primary diversity of bivalves in this part of the basin was not high: *Arcidae* and *Lucinidae* were absent, *Carditidae* were represented by the single genus *Scalaricardita, Veneridae* were rare, and species of subtropical genera were of minor importance (3 - 5%). This fauna displays a minor inheritance from the local late Eocene fauna: only 7%, of the Chegan species were found in the Ashcheairyk assemblages. The total number of the species inherited from the late Eocene constitutes 30%; however, similar to the assemblages from the Black Sea region and Ciscaucasia, these species are identical to the Latdorfian ones of northwest Europe. This fauna exhibits a small degree of endemism: about 15% (14 species) are not known outside the Transcaspian region.

Assemblages from the lower Oligocene deposits of Mangyshlak are less diverse (about 60 species) due to the predominantly clayey and deeper water enclosing deposits. The lower part of the section is more completely characterized by fossils that comprise a larger number of species inherited from the Eocene as well as members of some Eocene genera that did not occur in these regions during the Oligocene (*Eucrassatella, Aralocardia,* and *Pteria*). For the same reason, the share of subtropical elements is greater (about 15%). The endemism of the assemblages is minor (10%). The number of species identical to the northern Ustyurt mollusks is 22.

The early Oligocene molluscan assemblages from the southern Aral region, Kyzyl-Kumy, and Kopetdagh are less diverse and do not provide any additional data. There are almost no forms that indicate that the assemblages had a more thermophilic character than the Mangyshlak and Ustyurt ones. The rare exceptions are *Crassatina woodi,* present in the assemblages from the southern Aral region and *Glycymeris obovata,* from Kopetdagh. Endemic forms are almost completely absent as well.

During the second half of the early Oligocene, endemic brackish-water fauna spread throughout the Transcaspian region as well as the entire East Paratethys. It was much less diverse in the northeastern part of the East Paratethys than in the southern (Transcaucasian) one.

COMPARISON OF THE ASSEMBLAGES AND ANALYSIS OF THE CHARACTERISTIC AREAL DISTRIBUTION OF GENERA AND SPECIES

Tables 1 and 2 present a comparison of the discussed molluscan assemblages in pairs at the levels of genera-subgenera and species. The comparison shows a similarity between the faunas from West, Northwest, and East Europe and the faunas from the Transcaspian region. It also shows a distinction between these faunas and those of the Tethyan assemblages. The similarity of the South European assemblages, which was distinct in the late Eocene, can only be observed at the specific level in the early Oligocene, partly because the records are incomplete. The Preston difference coefficient shows that the East Paratethyan assemblages are distinctly uniform both at specific and generic levels.

The distribution areas of individual species and genera appear to be wider in the Oligocene than in the late Eocene. Most genera and some species occurred throughout all of West Eurasia; the most widespread

Table 1. Comparison of the species composition of the Rupelian bivalves from West Eurasia

Preston coefficients

Assemblages	1	2	3	4	5	6	7	8	9	10	11	12
1. N. Italy	**85**			0.79	0.91							
2. S. Bulgaria		**36**	0.83	0.83		0.91	0.91	0.93	0.97	0.93	0.94	0.95
3. Armenia		11	**76**	0.85			0.99	0.97				
4. Romania	12	7	8	**28**		0.96	0.98	0.96	1.00		0.95	0.98
5. Aquitaine	12				**137**	0.89	(0.97)					
6. C. France		6		4	18	**110**	0.78	0.89	0.93	0.86	0.86	0.87
7. Bavaria		7	3	2	(5)	33	**122**	0.74		0.79	0.83	0.86
8. Belgium		4	3	1		10	21	**45**	0.82	0.85	0.81	0.84
9. Germany		1		0		5		10	**41**	0.78	0.71	0.75
10. Georgia		6	8	5		18	29	12	16	**106**	0.66	0.65
11. Black Sea region		4		2		18	20	14	19	36	**74**	0.54
12. Ustyurt		4		2		17	19	13	19	44	45	**96**

Number of common species — Number of species

Notes to Tables 1 and 2: Semibold numbers placed diagonally indicate species in the assemblages (genera in Table 2); high Preston coefficients values are hatched; the coefficients showing the closest connection are doubly hatched (for threshold values see Popov *et al.*, 1993; Popov, 1994).

Table 2. Comparison of the generic composition of the Rupelian bivalves from West Eurasia

Preston coefficients

Assemblages	1	2	3	4	5	6	7	8	9	10	11	12
1. N. Italy	**41**	0.91	0.85	0.77	0.79	0.83	0.81	0.87	0.79	0.82	0.81	0.83
2. S. Bulgaria	4	**24**	0.69	0.91	0.72	0.78	0.72	0.93	0.92	0.87	0.86	0.83
3. Armenia	9	13	**54**	0.74	0.64	0.82	0.78	0.83	0.79	0.77	0.75	0.74
4. Romania	9	8	12	**23**	0.81	0.78	0.72	0.93	0.92	0.87	0.86	0.83
5. Aquitaine	16	14	19	10	**97**	0.64	0.63	0.79	0.83	0.74	0.82	0.77
6. C. France	11	10	14	12	34	**64**	0.58	0.73	0.75	0.61	0.72	0.68
7. Bavaria	14	13	18	11	39	35	**78**	0.57	0.67	0.60	0.63	0.63
8. Belgium	6	3	10	5	16	20	27	**40**	0.65	0.68	0.58	0.65
9. Germany	9	3	10	3	11	13	17	14	**29**	0.53	0.54	0.73
10. Georgia	12	7	18	11	28	32	37	20	22	**73**	0.46	0.37
11. Black Sea region	12	6	17	10	16	21	29	23	20	38	**53**	0.33
12. Ustyurt	11	8	19	11	22	26	32	21	21	49	43	**66**

Number of common genera — Number of genera

forms were *Barbatia nysti, Glycymeris lunulata, Pycnodonte gigantica callifera, Glossus subtransversus, Callista splendida, Pelecyora westendorpi*, and *Corbula gibba*. Some species with broad distributions were not found in the northeastern part of the East Paratethys, probably because the water environments here were the coldest. These species are *Glycymeris obovata, Anadara sulcicosta, Chlamys hoeninghausi, Claibornites tenuistria*, and *Ctena squamosa*.

As in the late Eocene, the early Oligocene was characterized by the most thermophilic genera and species with a predominantly Tethyan distribution (*Crassatella* s.s., *Pinna, Lima* s.s., *Pseudomiltha, Trachycardium, Discors, Chlamys biarritzensis, Spondylus cisalpinus,*

Table 3. Distribution of bivalve genera and subgenera in the early Oligocene basins of West Eurasia

Genera and subgenera \ Geographic distribution	N. Italy	Bulgaria	Armenia	Romania	Aquitaine	C. France	Switzerland	Bavaria	Belgium	N. Germany	Poland	Georgia	Black Sea region	Mangyshlak + Ustyurt
	1	2	3	4	5	6	7	8	9	10	11	12	13	14
I. Genera of the Tethyan distribution														
Trisidos	1													
Propeamussium s.s.			1											
Chlamys (*Peplum*)					1									
Dimya			1											
Lima s.s.		1			2									
Pseudomiltha	2			1	1									
Boeuvia		1	1											
Divalinga	1		1											
Fimbria	3													
Microstagon					1									
Paraglans					1									
Venericardia			1											
Pteromeris					2									
Crassatella	8	2	4		1									
Crassinella					1									
Cardium s.s.	1													
Trachycardium	1			1	1									
Vepricardium	2													
Corculum	1													
Loxocardium			1		2									
Granocardium			1											
Discors	1		1	1	1									
Avicularium		1												
Bissocardium					1									
Goniocardium				1										
Tresus				1										
Ensiculus					1									
Macrosolen		1		1										
Glossus (*Meiocardia*)					1									
Periglypta	1		1		1									
Meroena					1									
Callocardia (*Atopodonta*)					1									
II. Genera in common between the Tethys and Central Europe														
Arca	1	1			1	1	1	2						
Acar					1			2				1		

Table 3. (Contd.)

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Barbatia (*Obliquarca*)			1		1									
Glycymeris (*Tucetilla*)	1				2	1		1	1					
Lithophaga		1			2		1	1						
Brachydontes					1			1						
Septifer	1				1			1						
Atrina			1		1	1		1				1		
Pinctada					1			1						
Isognomon					1	1		2	1			2		
Propeamussium (*Parvamussium*)		1	1		1			1						
Palliolum s.s.								1						
Chlamys (*Peplum*)					1									
Ch. (*Flexopecten*)								2						
Plicatula					2			1						
Spondylus		4	3	2	2	1	1	1	1					
Limaria								1						
Crassostrea					1	1	1	1						
Lucina s.s.												1		
Linga s.s.					1	1								
L. (*Bellucina*)		1												
Cavilucina						1								
Parvilucina					1									
Claibornites		1				1	2	2	1			1		
Ctena					1	1	1	1				1		
Anodontia	1		1	1		1	1					?		
Megaxinus s.s.			1		2							1		
M. (*Gibbolucina*)	2		1		1									
Cardita s.s.					1									
Megacardita					2									
Europicardium	2		1									1		
Peronaea	1													
Arcopagia					1		1	2	1					
Donax	1				1	1								
Asaphis						1								
Lutetia					1	1								
Coralliophaga					1									
Timoclea					1			2						
Tivelina					2			1						
Clausinella					1									
Circe					2									
Petricola (*Rupellaria*)					1									
Corbula (*Ficusocorbula*)			1											
Gastrochaena					2	1		1						
Kuphus					1									

Table 3. (Contd.)

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
III. Widespread genera														
"Nucula"					1	1	2		1	2	1			
Nucula s.s.								2						
N. (Lamellinucula)			2				1	3	1	1	3	4	4	3
Leionucula			1								1	1		1
Tindaria												1		
Nuculana								1		2	1	1	1	
N. (Saccella)			2			1	1		1	1	3		1	2
N. (Jupiteria)			1						1		1	1	2	2
N. (Ledella)												1		1
Solemya			1											
Barbatia s.s.	10	3	3		4		2	2		1	3		1	1
Anadara	2							1				1	1	
Bathyarca			3									1	1	1
Arcopsis					1	1	1	1			1			
Trinacria									1					
Limopsis s.s.						1		1	1	1		1		
L. (Pectunculina)					1			1	1	1	1		2	1
Glycymeris s.s.	4	1				2	2	1	1	1	1	1	1	1
Mytilus						1				1				1
Arcomytilus												1		2
Arcoperna					1		1	1				1		1
Crenella					1	1								
Semimodiola					1									
Musculus					1	2		1				2		
Modiolus					2	3				1		2		1
Pteria			1		1	1		2				1		1
Lentipecten		1	1	2							1		1	1
Palliolum (Delectopecten)			1					1					1	1
"Chlamys"										1		3	1	
Chlamys s.s.		1	2	1	3	3	2	5						1
Ch. (Aequipecten)		3	2	1	4			2	1			1	3	3
Ch. (Hilberia)								2	1		1			
Pecten		1	1		1					1		2		1
P. (Flabellipecten)													1	1
Anomia			1	1	1			6		3		2	1	1
Limatula						1		1				1		1
"Ostrea"					2					1		1		
Cubitostrea		1	1	2	2	1	1							
Pycnodonte		3	3		2	1	1	1	1					
Crassostrea												1		
"Lucina"	3				5	2	3			1				
Megaxinus (Saxolucina)			2			1	1	3	1			2	2	2

Table 3. (Contd.)

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Lucinoma			1											
Loripes					1									
Thyasira			3						1	2	2	2	2	2
Diplodonta					1	3		1	1			1		
Chama	7		1	1			1	1		1				
Aligena					1									
Kellia					3									
Erycina						5								
Spaniorinus					1				1					
Scintilla						1								
"*Cardita*"						1								
Acturellina												1	1	
Cardites												2	2	3
Carditopsis					1									
Astarte	1							7	3	5	1	2	3	4
Crassatina			1			1		1					1	
Eucrassatella													1	1
"*Cardium*"						2								
Laevicardium												1		
Parvicardium	1				1	1	2	3	1			2	1	3
Plagiocardium	1				1									
Acanthocardia			1											
Nemocardium s.s.	2					1			1			1	1	1
N. (*Habecardium*)	4		1	1		1		2	1	1	1	2	2	2
Mactra				1		1								
Spisula												1		1
Lutraria					1							1		1
Ervilia					1									
Solen			1	1										
Cultellus					1	1						1		2
Ensis														1
Siliqua			1			2						2	1	1
"*Tellina*"	3			1		6							1	
Moerella					1							1	1	
Angulus			1			3	1	1			1	1	1	1
Peronidia												1		
Gari s.s.	1				1		1	1				2	1	1
G. (*Psammodonax*)					1									
Psammobia						3	1	1	1					
Sanguinolaria									2					
Abra					1	6	1	1	1	1	1	3	2	3
Congeria					1								1	
Arctica							1	1	1	1	1	2	1	1

Table 3. (Contd.)

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Trapezium														1
Glossus			1			1		1	1	1		2	1	1
Polymesoda		2		1		2		2				1	2	2
"*Venus*"		1				1								
Pitar			4		1				1				3	
Callista s.s.	1			2	3	8								
C. (*Nitidavenus*)	1							1				1		
Chione			2	1										
Pelecyora (*Cordiopsis*)		1		1	1	1	1	1	1			1	1	1
Sphenia					1	4			1			1	1	4
"*Corbula*"					2	3								
C. (*Caryocorbula*)	1			1			1	1		1		1	1	1
C. (*Variocorbula*)					2	1	1	1			1	2	1	2
Caestocorbula						1	1	2	1					
Lentidium				1	1	3	1					1	1	4
Spheniopsis								1	1					
Hiatella					1	1		2	1					
Panopea	9			1		1	1	1	1			2		
Panomya								1						
Martesia						1	1							
Jouannetia					1	2								
Parapholas									1					
Pholadidea					1									1
Xylophaga								1						
Teredo							1	1		1				
Pholadomya s.s.		1	1											
P. (*Bucardiomya*)	1				1		1	1				4	2	1
Lyonsia														1
Thracia					1	1	2	1	1		1	3	1	1
Poromya						2								
Cuspidaria						1						1	2	1
Cardiomya										1		1		1
Clavagella					1		2	1						
IV. Genera absent from the Tethyan basin														
Portlandia								1						
Yoldiella								1				1	1	1
Yoldia					1									
Nucinella								1	1				1	
Nucunella									1					
Dacridium													1	1
Gonimyrtea										1		1	1	1
Scalaricardita						1	2	1	2	2	2	3	3	2
Cerastoderma							1					1		1

Table 3. (Contd.)

	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Pygocardia								1			1		1	1
Dosiniopsis									1				1	1
Janschinella												1		
Corbula (*Lenticorbula*)														1
Cyrtodaria												1	1	1
Phragmorisma							1	1				1		1
Number of species	85		76		137		55		50		30		74	
		36		28		110		122		41		106		96
Number of genera and subgenera	41		54		97		44		45		22		83	
		24		23		64		78		29		73		66

Periglypta aglaurae, and *Macrosolen hollowaysi*) (Fig. 2A). A great number of genera (*Yoldiella, Nucinella, Scalaricardita, Pygocardia, Phragmorisma,* and others; group IV in Table 3) and species (the most widespread forms are *Nucula chasteli, Astarte gracilis, Pholadomya weissi, Cyrtodaria angusta,* and others; Fig. 2A) were not found in the most thermophilic assemblages from southern Europe, Aquitaine, southern Bulgaria, Transylvania, and Armenia but were abundant in the northern faunas.

Another characteristic group was the thermophilic genera and species (*Arca biangula, Atrina, Isognomon, Spondylus, Claibornites, Ctena,* and others; Fig. 2B), which inhabited the West and Central European basins and the southern (Transcaucasian) part of the East Paratethys and did not occur in the North Sea basin (except for Belgium) and most of the East Paratethys. The distribution of these subtropical forms was probably limited by climatic factors. In contrast, *Pecten stettinensis, Nemocardium excomatulum,* and *Dosiniopsis sublaevigata* were spread to the north of the Tethys. The great number of species limited in distribution by the East Paratethys should be mentioned as well (20% of this fauna). Among them, there are abundant species spread widely in the East Paratethys, such as *Yoldiella chadumica, Pterolucina batalpaschinica, Cardites kenderlensis,* and others.

In the second half of the early Oligocene, after the closing and freshening of the Paratethys, the first common endemic species and genera appeared between the West European and East Paratethyan basins. These are *Ergenica, Jansehinella, Corbula* (*Lentieorbula*), and *Cerastoderma serogosicum.*

ZOOGEOGRAPHIC DIVISION

The above data implies that the sharp distinctions between the late Eocene faunas of the southern (African) and northern (European) coasts of the Tethys Sea were smoothed in the early Oligocene owing to faunal exchange and the fact that the African coast became a part of the Mediterranean region (Fig. 3).

In the early Oligocene, the differences between the Mediterranean and North European regions still remained, but the boundary between them was displaced southward. As in the Priabonian, it lay between the Aquitainian and Parisian basins of West Europe. Most of the Central Europe basins were inhabited by fauna of the North European type, although some thermophilic mollusks undoubtedly of Tethyan origin are known from the Transylvanian basin. In the Oligocene, a sharp boundary occurred that was probably caused by the continental barrier within the Transcaucasian basin, which was the only one in the early Priabonian. Southern Armenia was inhabited by Tethyan fauna, while southern Georgia became a part of the East Paratethys, which belonged to the North European region.

A characteristic feature of the molluscan assemblages of the Mediterranean region of the Tethys was that the tropical and subtropical genera accounted for 70%. To a considerable extent, this fauna was inherited from the Eocene: the number of new species did not exceed 30%, and many genera and species did not reduce their distribution and remained inherent to the ancient Mediterranean faunas. About 30 bivalve genera peculiar to the fauna of the Mediterranean region made it distinguishable from the assemblages of the northern areas; the Preston difference coefficients are 0.74 - 0.93 at generic level and 0.91 - 1.0 at species level. However, since the late Eocene, the fauna of the Mediterranean tropical-subtropical region underwent some considerable changes: (1) It lost its uniformity, and all of the discussed assemblages greatly differed from each other. This can be attributed to their insufficient representativeness against a background of very diverse primary fauna. (2) Both generic and species compositions were noticeably impoverished. (3) Species of North European origin appeared in every Mediterranean assemblage. Thus, *Barbatia nysti, Anadara sulcicista, Glycymeris lunulata, G. obovata, Astarte kickxi,*

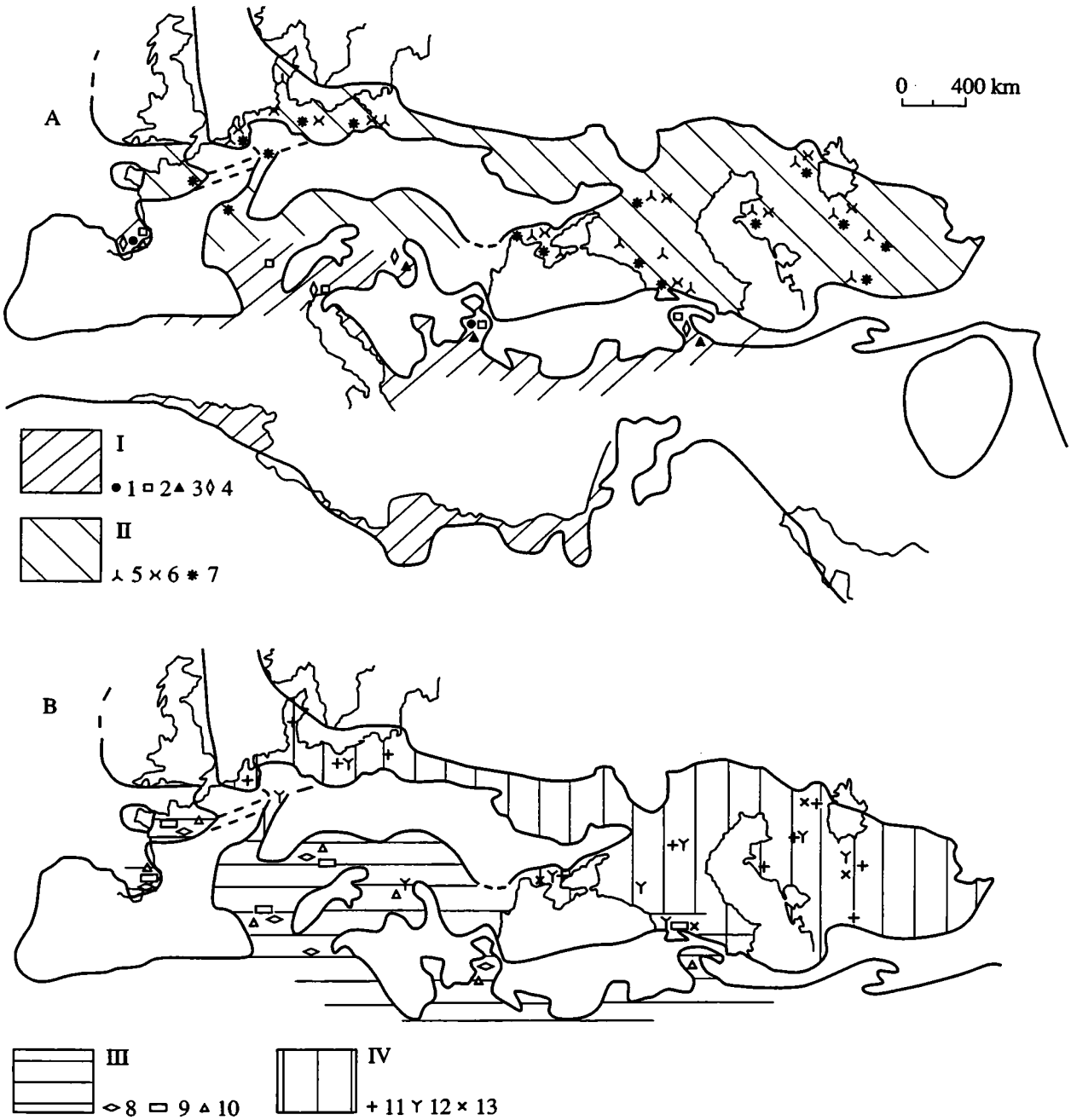

Fig. 2. Areas of distribution of characteristic Rupelian molluscan genera and species.
(A) I. Tethyan genera: *Lima* s.s., 2 – *Crassatella*, 3 – *Macrosolen*, 4 – *Discors*; II. North European genera and species: 5 – *Nucula comta*, 6 – *Nemocardium excomatulum*, 7 – *Scalaricardita*; (B) III. Some thermophilic genera: 8 – *Arca*, 9 – *Ctena*, 10 – *Spondylus*; IV. Cold-water genera and species: 11 – *Pecten stettinensis*, 12 – *Gonimyrtea*, 13 – *Cyrtodaria*.

Nemocardium tenuisulcatum, Glossus subtransversus, and *Pelecyora westendorpi* were widespread in the Mediterranean region during the early Oligocene.

In the Rupelian bivalve assemblage of the North European region, species of tropical genera were absent, tropical-subtropical genera constituted less than 50%, and some species of boreal genera occurred. The renewal of the molluscan composition was more noticeable, and Eocene species comprised less than 50% of the assemblage. Most of these species were identical to those of the Latdorfian fauna, which had transitional Eocene–Oligocene species compositions. The North European fauna was dominated by genera with a broad distribution. However, it was distinguished by nearly 15 genera and subgenera from the fauna of the Mediterranean region. The faunal unifor-

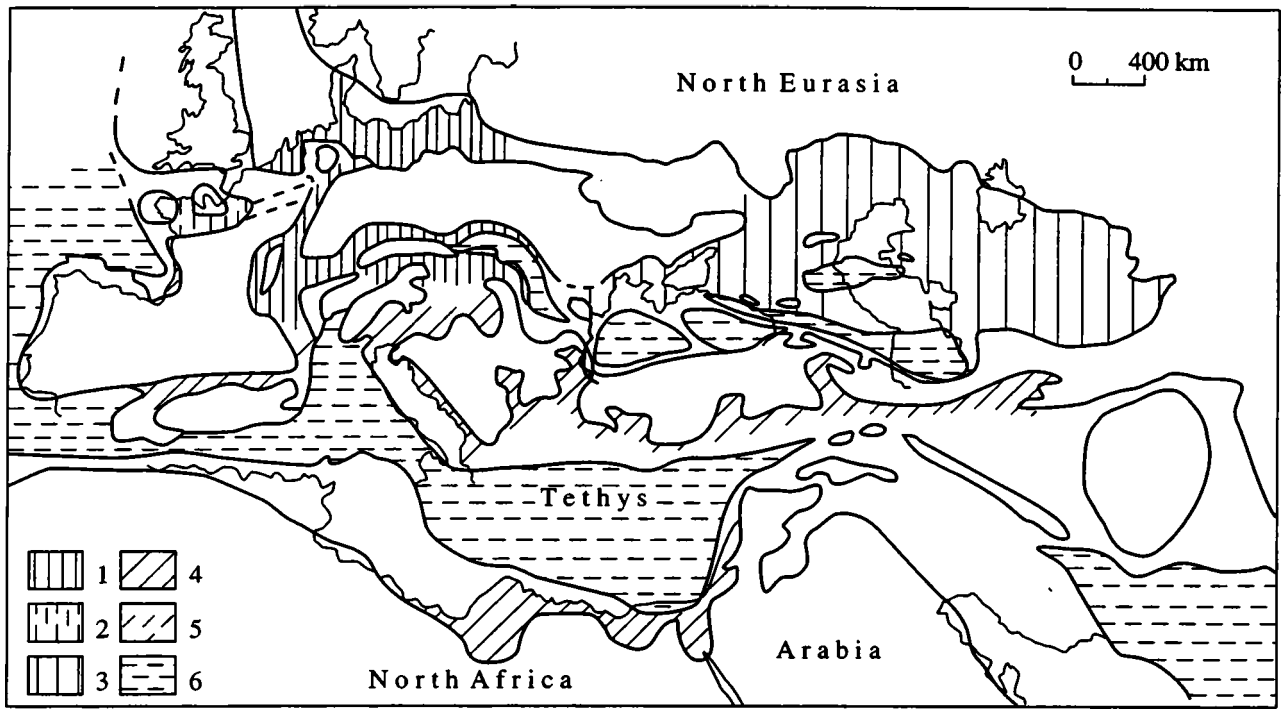

Fig. 3. Zoogeographic division of the West Eurasian sea shelves based on the early Oligocene bivalve records.
1 - 3 – North European region: 1 – German Province, 2 – Anglo-Parisian Province, 3 – East Paratethyan Province; 4 – Mediterranean region; 5 – Aquitaine Province; 6 – the extrashelf deep-water areas of the basins.

mity of the North European region is apparent at the generic and species levels, but relatively abundant data on the Rupelian fauna allows the region to be subdivided into smaller zoogeographic units. According to climatic peculiarities, this region can be considered as a subtropical-boreal one, which comprises a warm-water southern subtropical zone and a cool-water northern zone with predominantly boreal fauna. The fauna of the first zone includes the molluscan assemblages from the Parisian and Belgian basins, the Mainz, and many other basins of Central Europe, as well as from the southern (Transcaucasian) part of the East Paratethys. The assemblages of the northern and eastern parts of the East Paratethys and, probably, those of the North German basin are referred to the second zone. The cool-water aspect of the North German assemblages was probably determined by the lack of shallow-water assemblages.

However, the above zonality does not correspond to the units of provincial rank. Thus, the distinct uniformity of the East Paratethyan fauna allows this area to be recognized as a single East Paratethyan province, despite the climatic difference between the northern and southern parts. This uniformity is evidenced not only by mollusks, but by benthic foraminifers as well: the Transcaucasian foraminiferal assemblages are to those of the northern areas. Species endemism accounted for 20% for this province. The faunal distinctions between the northern and southern (Transcaucasian) parts of the East Paratethys should probably be considered as subprovincial.

Some other provinces of the North European region can also be outlined. An original and independent early Oligocene fauna existed in the Parisian basin. The composition of this fauna and the more diverse and thermophilic fauna of the Aquitainian basin shows the zoogeographic independence of the Atlantic mollusks.

Similar molluscan assemblages of the North Alpine and Mainz basins differed from those of the Norwegian-Greenland Sea. However, the transitional aspect of the fauna from the Belgian basin suggests that these differences were due to the facies environments. The faunas of all these basins should probably be regarded as belonging to a single German province, which consisted of northern and southern subprovinces (Fig. 3).

There are contradictory data provided by various faunal groups concerning the degree of faunal independence of the freshwater Paratethyan basin during the Solenovian (late Rupelian) time. According to bivalve records, it was very high. If the brackish-water fauna of the first stage of development of this basin is considered alone, of the 10 genera and 18 species that inhabited the Protoparatethys, only 6 genera (*Janschinella, Ergenica, Urbnisia, Rzehakia, Merklinicardium,* and *Korobkoviella*) and 11 species were reported from this basin. This high degree of endemism is comparable with that of the recent Caspian fauna, which belongs to a zoogeographic unit of high rank (an independent

realm; Starobogatov, 1970). However, the Solenovian gastropods were dominated by widespread lagoon, brackish-water, and freshwater forms, similar to those from the Oligocene basins of Europe. Only one species was endemic. In the ostracod fauna, 75% of the species were of marine origin, which were widespread in the Rupelian basins, and some eurybiontic forms and a few endemics occurred (Popov *et al.*, 1985). These data imply that, in the late Rupelian, the Paratethys was an independent faunal province (the Protoparatethyan province, according to Rusu, 1988), which belonged to the North European region and comprised two subprovinces: northern and southern (Transcaucasian) ones. The Central European basins were referred to the northern subprovince, due to the occurrence of *Ergenica* and *Janschinella garetzkii* and the absence of various Cardiidae and other Transcaucasian species. If these inferences are confirmed by further investigations (the records of the Central European fauna are still sparse), the subprovincial division will appear to be controlled by other factors, probably including the degree of salinity, rather than climatic conditions, as it was in the Oligocene sea basins.

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Reviewer M. A. Akhmet'ev

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Palynostratigraphy of Upper Quaternary Deposits of Sverdrup Island, the Kara Sea

P. E. Tarasov*, A. A. Andreev, F. A. Romanenko*, and L. D. Sulerzhitskii*****

** Moscow State University, Moscow, 119899 Russia*
*** Institute of Geography, Russian Academy of Sciences, Staromonetnyi per. 29, Moscow, 109017 Russia*
**** Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia*

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Abstract – New palynologic data and radiocarbon datings, which were obtained for a buried peat-bog from Sverdrup Island, allowed the landscape-climatic environments to be reconstructed for this region for the period 9.5 - 12 ka ago. A new approach to the problem of Late Quaternary glaciation on the Kara Sea shelf was also made feasible. Three stages were distinguished in the evolution of the vegetation and climate: a relatively warm and humid climate corresponding to the Allerød, with tundra-type vegetation; a cold and arid climate corresponding to the late Dryas, with dominating xerophytes; and a stage of the warmest climate corresponding to the beginning of the Preboreal interval of the Holocene, with dominating trees and bushes. The available data indicated the absence of a late Pleistocene (Sartanian) ice sheet on the Kara Sea shelf.

Key words: *Sartan Ice Age, shelf, palynostratigraphy, late Pleistocene, Holocene, radiocarbon dating.*

Palynologic data from high-latitude areas of Siberia and, especially, data from the paleobotanic analysis of peat-bogs related to the end of the late glacial period and the beginning of the early Holocene are not numerous (Makeev and Ponomareva, 1988; Ukraintseva *et al.*, 1989; Makeev *et al.*, 1992) but are of great significance for landscape and climatic reconstructions. These data are particularly important when considering glaciation problems on the Arctic shelf in the late Pleistocene. Judging from the present depth of the Kara Sea shelf, the Svedrup Island was part of the continent during the late Pleistocene. The sea level was at –40 to –60 m 11 - 13 ka ago (Kaplin, 1975). Erratics as well as the exaration and accumulation landforms described for the Sverdrup and Kirov islands (Dibner and Zakharov, 1970) suggest that the islands may have experienced glaciation in the past. Some researchers have assigned this glaciation to the Late Quaternary (Sartanian) (*Geomorfologicheskoe Raionirovanie* ..., 1980; Grosval'd, 1983). According to other data (Danilov, 1987; Danilov and Polyakova, 1989; Makeev *et al.*, 1992), shelf areas were free of the late Pleistocene ice sheet. New polynologic data as well as radiocarbon datings, which were obtained for a buried peat-bog on Sverdrup Island, allow the landscape-climatic conditions that existed in this area 9.5 - 12 ka ago to be reconstructed.

Romanenko (Geography Department, Moscow State University) conducted field studies and sampling in July - August 1992 during investigations being performed by the Arctic Expedition from the Severtsov Institute of the Evolutionary Morphology of Animals, Russian Academy of Sciences. Andreev (Institute of Geography, Russian Academy of Sciences) and Tarasov (Geography Department, Moscow State University) performed a palynological analysis and interpreted spore-and-pollen spectra; Sulerzhitskii (Institute of Geology, Russian Academy of Sciences) performed the radiocarbon datings. The botanic composition was analyzed in the Laboratory of the Geological Production Association *Torfgeologiya* with assistance from Uspenskaya. Kozhevnikov (Institute of Botany, Russian Academy of Sciences), a participant of the Arctic Expedition, submitted data on a recent flora.

Sverdrup Island (74°30′ N, 79°30′ E) is situated on the Kara Sea shallow-water shelf, 110 km off the coast (Fig. 1A), and belongs to the geomorphological subarea of the Yamalo-Gydansk bank. The sea depth in this region does not exceed 15 - 25 m, and a 50-m isobath approximates the island at a distance of only 150 km to the north.

The island was discovered on August 18, 1893, from the board of the famous vessel *Fram* and was named by F. Nansen in honor of the ship master. The first disembarkation was made in August 1933 by participants of expeditions on the ice breakers *Sibiryakov* and *Rusanov*. At that time, Gakkel and Vlodavets described the topography and geological structure of the island (Vlodavets, 1933).

The area of the island spans 70 km^2. It represents an accumulative terracing plain, highly dissected by a ravine-balka system (Fig. 1B). About one-third of its area is occupied by extensive sand banks, which are

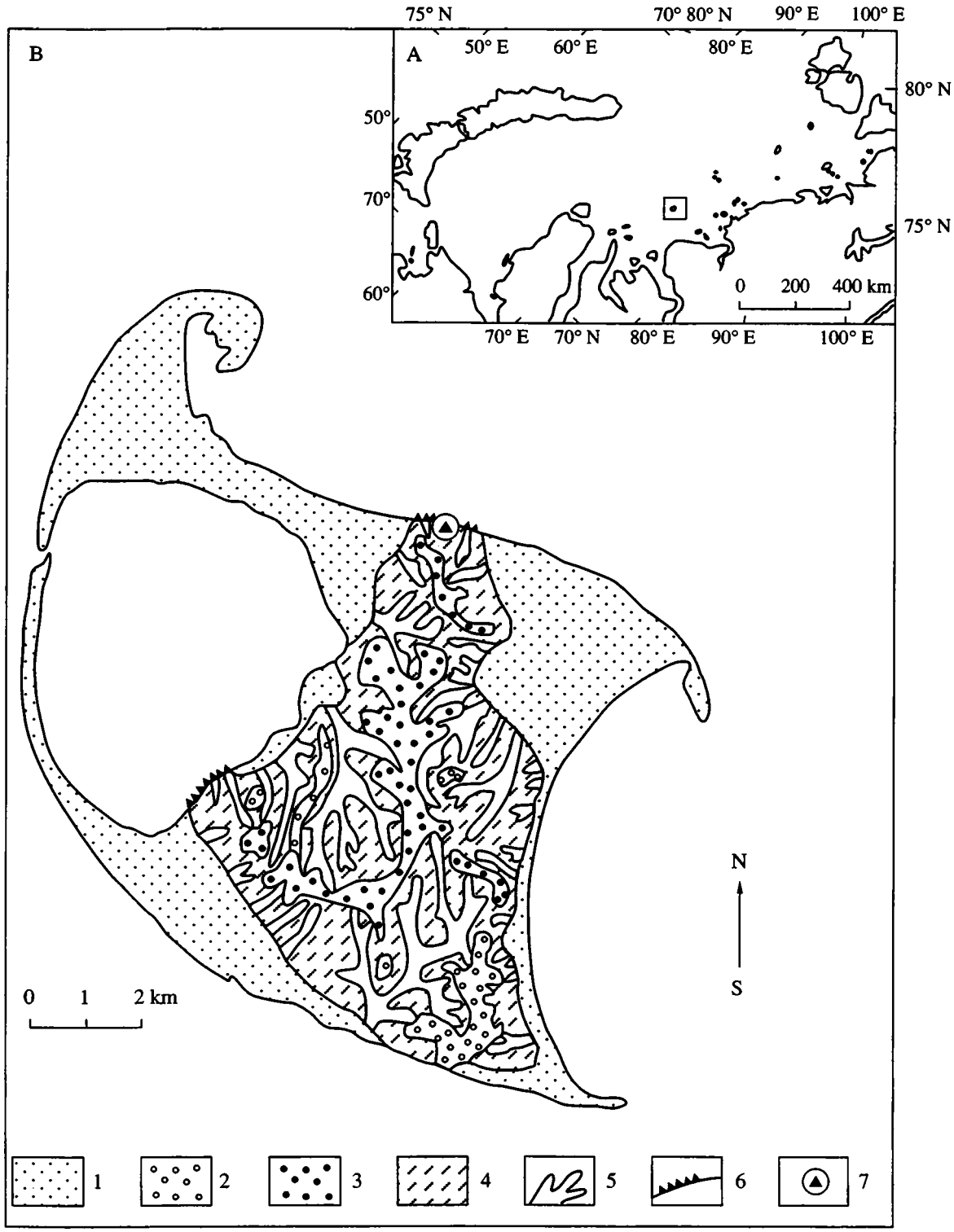

Fig. 1. Geographical position (A) and geomorphological scheme (B) of Sverdrup Island. 1 – modern marine terrace (0 - 2 m); 2 – terrace 12 - 18 m high; 3 – terrace 22 - 28 m high; 4 – slopes; 5 – valleys of temporary streams; 6 – scarps; 7 – location of the studied section.

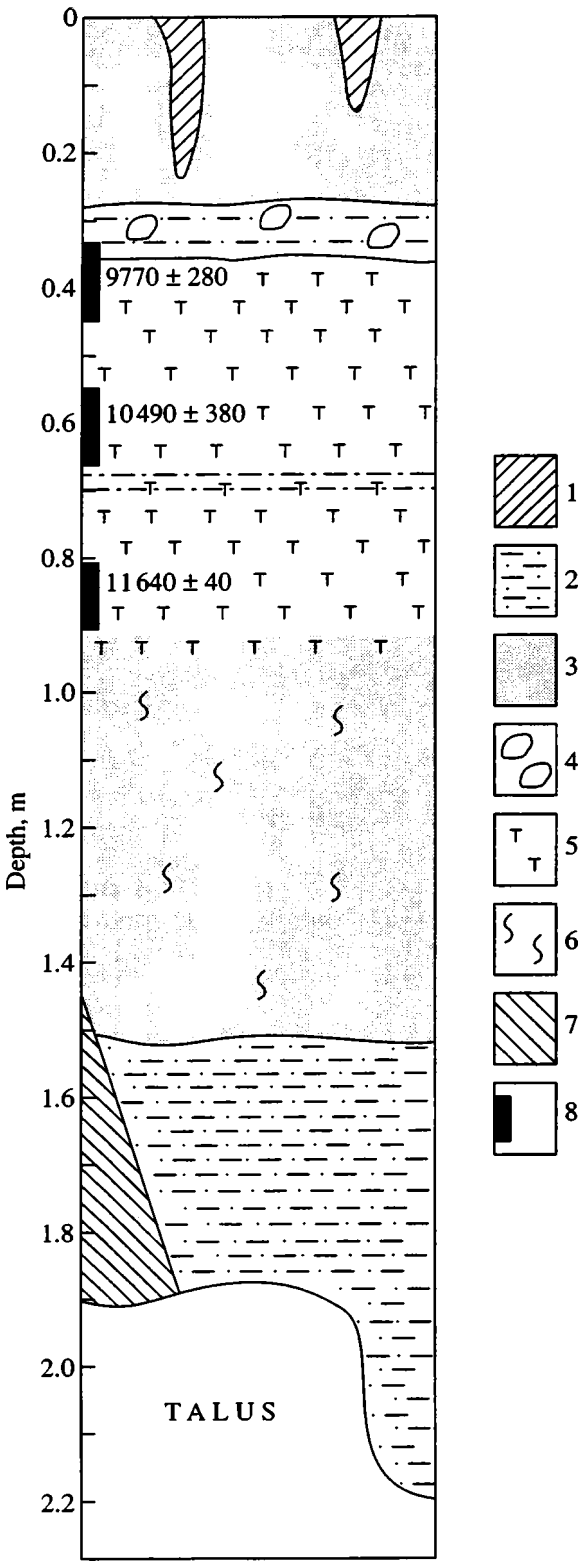

Fig. 2. Sequence of loose deposits on the northern shore of Sverdrup Island.

1 – loam; 2 – loamy sand; 3 – fine sand; 4 – pebbles; 5 – peat; 6 – plant detritus; 7 – ice wedges; 8 – sampling sites for ^{14}C dating.

as high as 2 m above sea level. The maximum height (33 m above sea level) occurs in the central part of the island. Country rocks are represented by Upper Cretaceous terrestial deposits overlapped by boulders and gritty-pebbly sands and loams of Quaternary age (Dibner and Zakharov, 1970; *Geomorfologicheskoe Raionirovanie* ..., 1980). A buried peat-bog is located in the northern part of the island, 7 - 8 m above sea level and at a distance of 6 m from the edge of the water.

The July temperatures on the island average 0.4 - 3.5°C, and the January temperatures are as low as –26 to –28°C; annual precipitation amounts to 200 - 300 mm. The duration of a frost-free period does not exceed 20 - 30 days (*Spravochnik* ..., 1967, 1969). The extremely severe climate here allows only herbaceous assemblages of Arctic deserts to exist. The vascular flora of the island is very poor: there are 34 taxons belonging to families Cyperaceae, Poaceae, Ranunculaceae, Juncaceae, Polygonaceae, Caryophyllaceae, Rosaceae, Saxifragaceae, Brassicaceae, Saliaceae, Papaveraceae, and Scrophulariaceae. Mosses and lichens are represented here by 27 and 33 species, respectively.

Samples for the palynological analysis were taken from an undistorted sequence, 2.2 m thick, at 10-cm intervals after clearing a cliff (Fig. 2). Further processing of the samples to separate the pollen and spores was performed according to the standard method (Grichuk and Zaklinskaya, 1948) in the Laboratory of the Latest Deposits and Pleistocene Paleogeography (Moscow State University). The palynologic spectra were studied using a Neovar-2 microscope with a magnification of 400. The peat samples appeared to be rich in pollen; whereas the samples taken from sand layers over- and underlying the peat, excluding those from a depth of 10 cm, contained no Quaternary palynomorphs. The number of redeposited paleospores in the underlying sands amounted to 20 grains per slide. Those samples that were very poor palynologically and contained many redeposited palynomorphs were excluded from further consideration. In the rest of the samples, the average number of calculated microfossils exceeded 200 grains. The pollen diagram (Fig. 3) was drawn with the help of the Tilia Graph Program. The percentage of palynospectrum components was calculated from the total spore–pollen quantity adopted as 100%.

Samples for radiocarbon analysis were collected from the peat part of the sequence (0.3 - 1.9 m) at intervals of 10 - 15 cm. Samples for the study of the botanical composition of the peat were taken from the same levels. Radiocarbon dates were determined by the method of counting on a liquid-scintillated apparatus, using benzol obtained from the radiocarbon of alkaline extracts (Arslanov, 1987) at the Laboratory of Isotope Geochemistry and Geochronology, Geological Institute, Russian Academy of Sciences.

A sequence of sands and loamy sands 1.3 m thick, with inclusions of ice wedges, was revealed at the base of the studied section (Fig. 2). These deposits contain

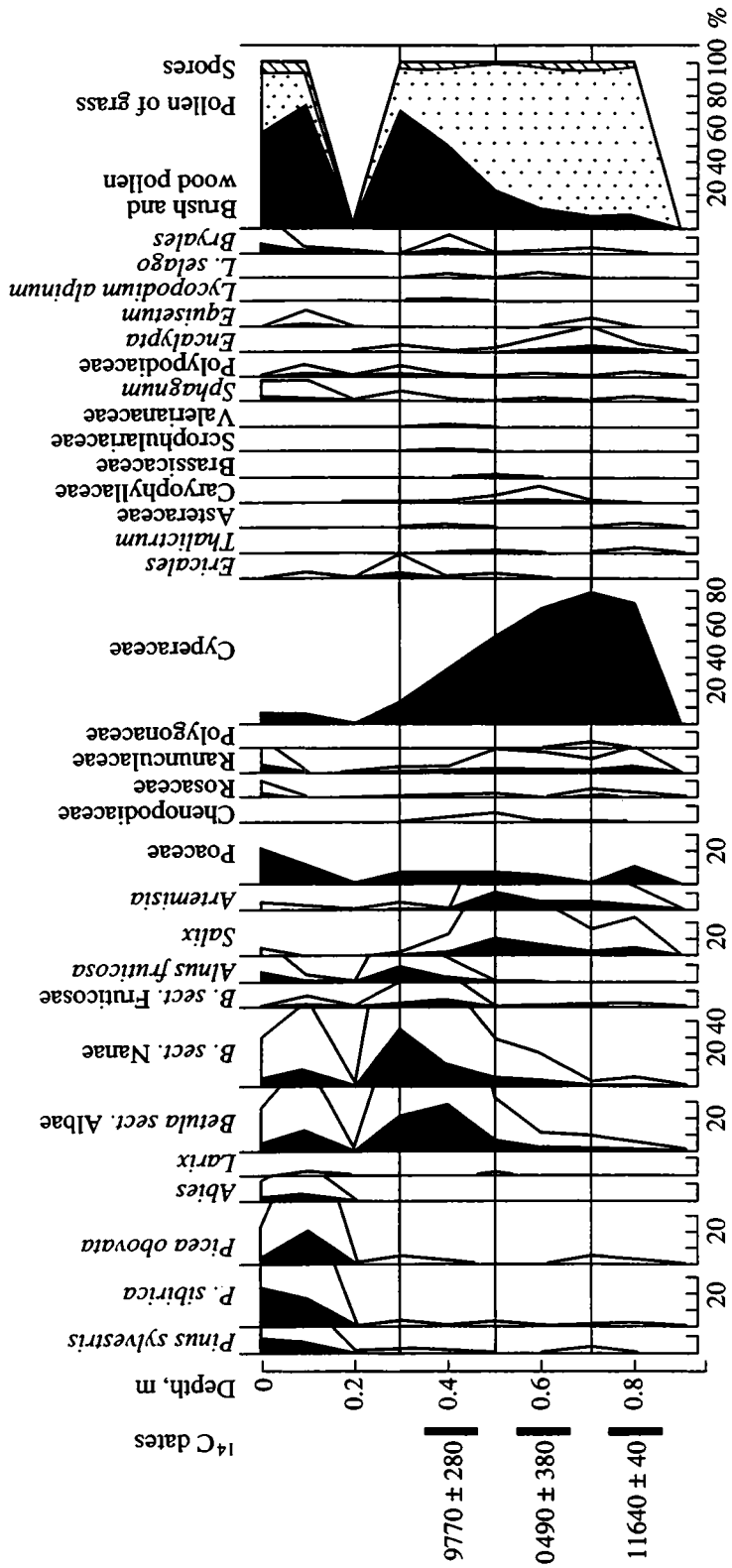

Fig. 3. Spore–pollen diagram of buried peat on Sverdrup Island.
The share of taxons with a low percentage is shown additionally with ×5 scale magnification.

very few Quaternary palynomorphs, which prevents any conclusion concerning the time of their accumulation. The relatively high content of redeposited Mesozoic spores indicates a genetic relation between the considered sequence and the Upper Cretaceous country rocks and, most likely, the eolian origin of the former. It may be suggested that the formation time of the ice wedges in the studied section corresponds to the late Pleistocene maximum cooling, when eolian and cryogenic processes were active in drained areas of the Kara Sea shelf under the conditions of an extremely continental climate. Such an assumption correlates well with conclusions made by other authors (Danilov, 1987; Danilov and Polyakova, 1989) concerning the absence of the late Pleistocene ice sheet in the northern part of western Siberia.

Upward the section, a horizon of buried peat (0.3 - 0.9 m) occurs, the lower layer of which, lying at a depth of 0.8 - 0.9 m, is as old as 11640 ± 40 years according to the radiocarbon dating (GIN-7625). Therefore, the accumulation of sedge eutrophic peat began in the Allerød (about 12 ka ago). The composition of the palynospectra suggests that the vegetation at that time was represented by steppe associations with *Artemisia*, Chenopodiaceae, and Asdteraceae along with assemblages including dominant representatives of Poaceae, Cyperaceae, *Salix*, Ranunculaceae, Rosaceae, *Polygonum*, Caryophyllaceae, *Valeriana*, and *Bryales*, which are typical for the modern tundra. Individual findings of pollen grains of spruce, pine, arboreal, and shrub birch indicate that these species are of eolian origin. The content of pollen in these arboreal species is several times lower than that in the subfossil spectrum, which indicates that their geographic range was much more remote than it is currently. Although birches of sect. Nanae grew on the peat surface, which is confirmed by findings of macroremains, they were low-yielding in pollen. The Allerød climate was drier than the modern one; the summer temperatures exceeded those of the present, whereas, conversely, the winter temperatures were much lower. The process of peat accumulation, which began in the Allerød, as well as the character of the related palynological spectra clearly indicate that the climate became milder, and the Sverdrup Island was free of an ice sheet at the end of late Ice Age.

The late Dryas cooling greatly affected the character of the pollen spectra, in which the share of *Artemisia*, Chenopodiaceae, and *Salix* increased considerably. The peat composition also changed, the botanical analysis of which showed a prevalence of green mosses among bog plants. According to radiocarbon dating, the age of the peat horizon from the depth of 0.55 - 0.66 m is 10 490 ± 380 years (GIN-7626). The expanding role of wormwood and mari species in the late Dryas vegetation indicates that the climate became substantially drier. The increased content of pollen of willows and birches of sect. Nanae as well as findings of their macroremains in the peat suggest an enchanced role of shrub coenoses on the peat surface, which also confirms that the conditions were less humid than in the Allerød.

The early Preboreal warming was marked by an abrupt change in the character of the pollen spectra. The pollen of trees and shrubs dominated, whereas that of steppe species practically disappeared from the spectra. The peat botanical composition did not change; only macroremains of birches vanished. This may be explained by the increased flooding of peat-bogs, which subsequently resulted in the disappearance of shrub coenoses from the peat-bog surface. Radiocarbon dates, which were obtained for the peat layer at a depth of 0.35 - 0.45 m as 9770 ± 280 years (GIN-7627) and for the underlying layers as 10 490 ± 380 years, suggest that the warming began about 10.2 - 10.3 ka ago, as it did throughout all of northern Eurasia (Khotinskii, 1977).

At that time, Sverdrup Island was characterized by vegetation typical of a modern southern tundra. Birches of sect. Nanae; Ericales; and, probably, *Alnus fruticosa*, which are absent in the modern flora of the island, dominated in the coenoses. It seems that arborescent birches did not to grow near the peat-bogs, but the high content of their pollen in the spectra (several times greater than that in the sample from the surface) undoubtedly indicates that birches of sect. Albae were distributed in the north at the beginning of Preboreal time.

The distribution of vegetation typical of a modern southern tundra, the pollen spectra, the complete disappearance of steppe associations, and the northern expansion of arboreal vegetation suggest that the climate of the first half of the Preboreal time interval was much milder than the modern one. Many researchers (Lozhkin, 1987; Makeev and Ponomareva, 1988; Ukraintseva, 1990; Nikolaev and Kolokolov, 1992; Ritchie *et al.*, 1983; Clague and Mathews, 1989) have reported that a climatic optimum existed in the early Holocene in high latitudes of the Arctic region. The principal reason for this warming, which appears to have been the maximum for the present Arctic maritime and insular regions, lies in the rapid sea-level rise by the beginning of Boreal time (Kaplin, 1975). This resulted in a transformation of regions with extreme continental climates into insular and maritime territories with corresponding climatic conditions. Therefore, the early Preboreal warming manifested itself on the Arctic coast as a thermal optimum under conditions of a climate that was not yet continental, although less severe than in the late Ice Age. The subsequent Subboreal warmings occurred under climatic conditions of the marine type due to the smoothing effect of cold waters from the Arctic Ocean. Even the maximum warming, which occurred 4.5 - 6.0 ka ago and was manifested as a climatic optimum in many regions of the northern hemisphere (Khotinskii, 1977), affected the Arctic nature to a lesser extent.

Peat accumulation ceased in the middle of the Preboreal time interval. Sands of eolian origin with vein pseudomorphs filled with loamy material, which overlay the peat-bog, contain very few pollen grains. This

prevents reliable interpretations of the character of landscape-climatic changes in the Holocene.

Samples from a depth of 10 cm and from the surface comprise pollen and spores that belong mainly to taxa of longdistant influx (*Pinus, Picea, Abies, Betula*, and *Alnus*) as well as to species of the local flora (Poaceae, Rosaceae, Cyperaceae, Ranunculaceae, *Salix, Bryales*, and *Sphagnum*). The pollen spectra composition and pollen ratio of longdistant and local species are those that are common for Arctic high-latitude areas (Krenke and Fedorova, 1961; Kalugina *et al.*, 1979; Van der Knaap, 1990; Andreev, 1992), reflecting the scanty vegetation of an Arctic desert.

CONCLUSION

A complex investigation of the buried peat-bog and the conducted paleoreconstructions allowed us to distinguish three evolutionary stages in the vegetation and climate during the late Ice Age and the beginning of the Holocene. The first stage corresponds to the Allerød and is characterized by a relatively warm and humid climate. The vegetation was represented by dominanting associations typical of a modern tundra; steppe coenoses also existed. The second stage had a cold and arid climate and is assigned to the late Dryas; it is characterized by an increasing role of xerophile vegetation. The third stage, which is the Holocene thermal optimum in this region (as in other Arctic maritime high-latitude areas), was attributed to the first half of the Preboreal time interval. It relates to a global warming at the boundary between the late Ice Age and the Holocene and to the relatively low level of the Arctic Ocean, which was responsible for the continental-type climate in the present maritime areas. The continental character of the climate allowed vegetation typical of a modern shrub tundra to grow.

Peat accumulation, which commenced about 12 ka ago during a global warming, ceased about 9.5 ka ago. This resulted from a rapid rise in the sea level (which caused a change in the climate from continental to marine) rather than from the late Preboreal cooling, which was recorded in many regions of northern Eurasia (Khotinskii, 1977). The reduction of the vegetative season and the considerable fall in summer temperatures negatively affected the flora composition of Sverdrup Island; birches of sect. Nanae, representatives of order Ericales, and *Alnus fruticosa* disappeared. The island vegetation cover acquired the character of the modern Arctic deserts.

The data obtained allow us to suggest that the ice sheet was absent in the northern part of western Siberia and on the Kara Sea shelf at the end of the late Pleistocene. Ice wedges preserved in sands and loams underlying peat-bogs and the scanty pollen revealed within them indicate that cryogenic and eolian processes played an active role in the vast areas of the drained shelf and the coastal lowlands in the late Quaternary.

Reviewer M.N. Alekseev

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BRIEF COMMUNICATIONS

New Data on the Geology of the Darbi Area, Western Mongolia

V. F. Shuvalov

Institute of Limnology, Russian Academy of Sciences, ul. Sevast'yanova 9, St.Petersburg, 196199 Russia

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The Darbi area (the Gobi Altai region of Mongolia) is important for geologic studies of the continental Mesozoic. A virtually complete Jurassic–Lower Cretaceous section with diverse faunal and floral remains is located south of the town of Darbi. This sequence has been visited repeatedly by many geologists and paleontologists (Devyatkin *et al.*, 1975; Neustrueva, 1982; Ponomarenko and Popov, 1980; Khosbayar, 1972, 1973; Shuvalov, 1982; *Yurskie ...*, 1985). Some of these authors noted that the Lower Cretaceous deposits outcropping south of the Gurvan Eren Range are unconformably overlain by Paleogene red beds. However, neither a description of the beds nor their organic remains have been reported to confirm this observation. In addition, no reliable evidence on the age of the beds underlying the Mesozoic deposits has been recorded. Moreover, since the first observation by Khosbayar (1972, 1973), the grey conglomerate member with a thickness of more than 300 m in the Baga–Khairhan Mt. area has been considered to be, as in other regions of Mongolia, a lower unit of the Lower-Middle Jurassic sequence. It should be noted, however, that organic remains were never recorded in these deposits. The only fossils recorded here were Lower-Middle Jurassic freshwater mollusks, which I found in 1970 in sand layers overlying conglomerates (Devyatkin *et al.*, 1975).

The higher Jurassic–Lower Cretaceous layers outcropping in the Gurvan-Eren Range and in its vicinity south of the Darbi locality contain abundant and diverse fossils (mollusks, conchostracan, dinosaurs, coprolites, plants, and others), which have provided a basis for reliable stratigraphic schemes of the Jurassic and Cretaceous in the studied region (Devyatkin *et al.*, 1975; Ponomarenko and Popov, 1980; Khosbayar, 1972, 1973; Shuvalov, 1982). I will not dwell here on the Mesozoic deposits of the Darbi area except for those sequences in the vicinity of the Baga–Khairhan Mt. and in the Guen–Khuduk area. In the former area, Mesozoic rocks overlay, and in the latter they underlay, respectively, the Paleozoic and Cenozoic deposits that are considered in this article (Fig. 1). The first plant and fauna remains from the Paleozoic and Cenozoic rocks, respectively, were sampled here during my field work in 1986. These finds either essentially changed or verified earlier ideas on the age of the rocks.

MIDDLE CARBONIFEROUS DEPOSITS

It should be noted that, due to their features, grey conglomerate deposits outcropping on the northwestern and western slopes of the Baga Khairhan Ula Mt. noticeably differ from the overlying grey and variegated predominantly sandy deposits, which are undoubtedly Early-Middle Jurassic in age. Earlier, these two members were considered as a single formation or complex (Devyakin *et al.*, 1975; Khosbayar, 1972). In the northern Baga Khairhan Mt. area, they are separated by an angular and azimuth unconformity (Fig. 1) and display different features. The conglomerate member resembles the underlying Paleozoic rocks, whereas the sandy member is more similar to the overlying Jurassic beds.

According to my observations (Fig. 2), the conglomerate member either conformably overlies the Lower-Middle Carboniferous volcanic rocks or, more frequently, has a tectonic contact with them (Fig. 1). The lava layers dip east-northeastward at an angle of 35° - 40°, whereas the conglomerate beds dip west-northwestward at an angle of 10° - 15°.

The conglomerate member is composed of the following sequence (from the base upward):

(1) Grey dense conglomerate with varying pebble sizes. The thickness is 10 m.

(2) Grey, dark-grey, and brown-grey coal-bearing siltstone, claystone, and sandstone with interlayers of dark-grey dense conglomerate and coal. The thickness of the sandstone and conglomerate beds is up to 0.5 m, whereas that of coal-bearing claystones and coal reaches 2 - 3 m and 1 - 1.5 m, respectively. The observed extention of lenslike coal-bearing beds exceeds 200 m. Claystone and sandstone above the coal beds yielded plant remains of *Abacanidium neuburge* (Dur.) (identifications by G.N.Vasil'eva). The thickness is 20 m.

(3) Grey and greenish-grey dense boulder to pebble conglomerate. The boulders and pebbles show a varying degree of roundness. Small clasts and lumps occur. The thickness is about 300 m.

To the south, the conglomerate member is unconformably overlain by grey and variegated sandstones with claystone, gravelstone, and conglomerate interbeds. These deposits (the Zhargalant Formation),

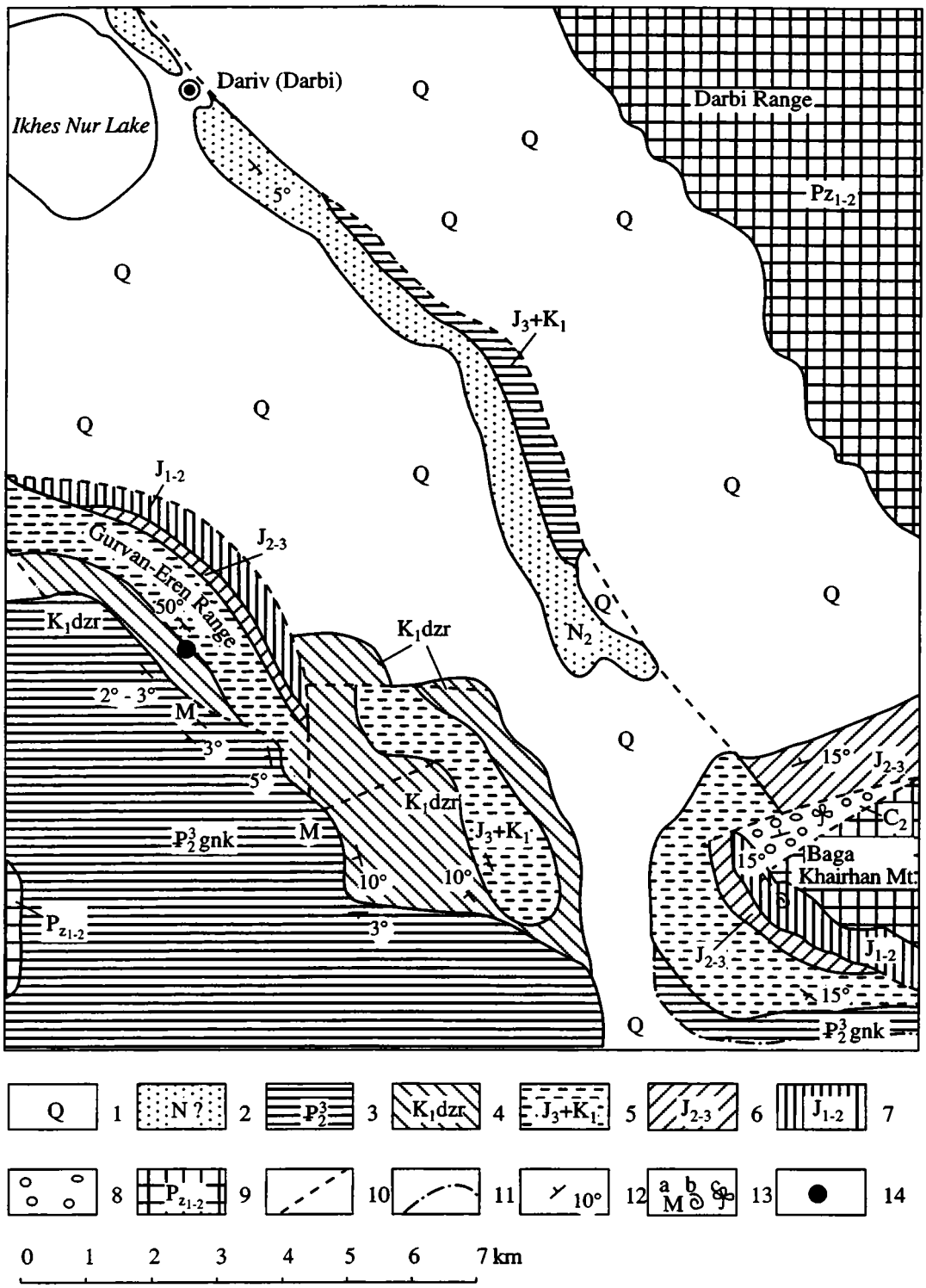

Fig. 1. Geological map of the Darbi area. Compiled by Shuvalov (1986).

1 – Quaternary; 2 – Neogene(?); 3 – middle Oligocene (the Guen Khuduk Formation); 4 – Lower Cretaceous, Aptian–Albian (the Dzereg Formation); 5 – Upper Jurassic–Lower Cretaceous; 6 – Middle-Upper Jurassic (the Darbi Formation); 7 – Lower-Middle Jurassic (the Zhargalant Formation); 8 – Middle Carboniferous (the Baga Khairhan Formation); 9 – Lower-Middle Paleozoic; 10 – tectonic contacts (faults); 11 – conventional boundaries; 12 – angle of layer dip; 13 – organic remains: a – mammalian, b – mollusks, c – leaf flora; 14 – Guen Khuduk Well location.

which are about 150 - 200 m thickness, contain the Early-Middle Jurassic mollusks *Pseudocardinia turfanensis* Mart., *P. jenisseica* Mart., and *Sibireconcha* sp. (Devyatkin *et al.*, 1975).

The sequence in the western vicinity of the Baga Khairhan Mt. is crowned by a member, about 100 m in thickness, of red sandstones, gravelstones, and conglomerates containing indeterminable dinosaur bones. These deposits correlate either with the Darbi Formation (Khosbayar, 1972) or with the upper part of the Middle Jurassic complex (Devyatkin *et al.*, 1975) and are conventionally dated as the Middle-Late Jurassic. In the western part of the mountain, they immediately overlie a conglomerate member that I consider to be the separate Baga Khairhan Formation.

In general, the lower conglomerate member, or the Baga Khairhan Formation, is grey in color. However, on the southern slope of the mountain, the conglomerates gradually change their color to pinkish-grey and even reddish-grey. Pebbles in the upper layers of the conglomerate member are smaller in size and have a higher roundness class than those in its lower part. Bed 2 of the described section is lenslike; its visible extension exceeds 300 m. The estimated total thickness of the Baga Khairhan Formation is 300 - 350 m.

The formation is dated by the remains of aforementioned *Abacanidium neuburge*, which, according to G.N. Vasil'eva (data after Durante, 1976), occurs in the Muruktsik Formation (the basal Middle Carboniferous) and correlative units of southern Mongolia. In western Mongolia, the deposits under consideration were recorded for the first time with similar flora.

Thus, the aforesaid implies that the conglomerate member differs sharply from the overlying Lower-Middle Jurassic deposits of the Zhargalant Formation as well as the underlying Paleozoic unit of highly metamorphosed volcanic rocks. It represents a separate stratigraphic unit that crowns the Paleozoic sequence here. The Mesozoic succession begins with the above-mentioned grey and variegated sandstones, conglomerates, gravelstones, and claystones (150 - 200 m in total thickness), which are exposed on the southern and southwestern slopes of the Baga Khairhan Mt. (the Zhargalant Formation) and contain mollusks of genera *Pseudocardinia* and *Sibireconcha* (Devyatkin *et al.*, 1975).

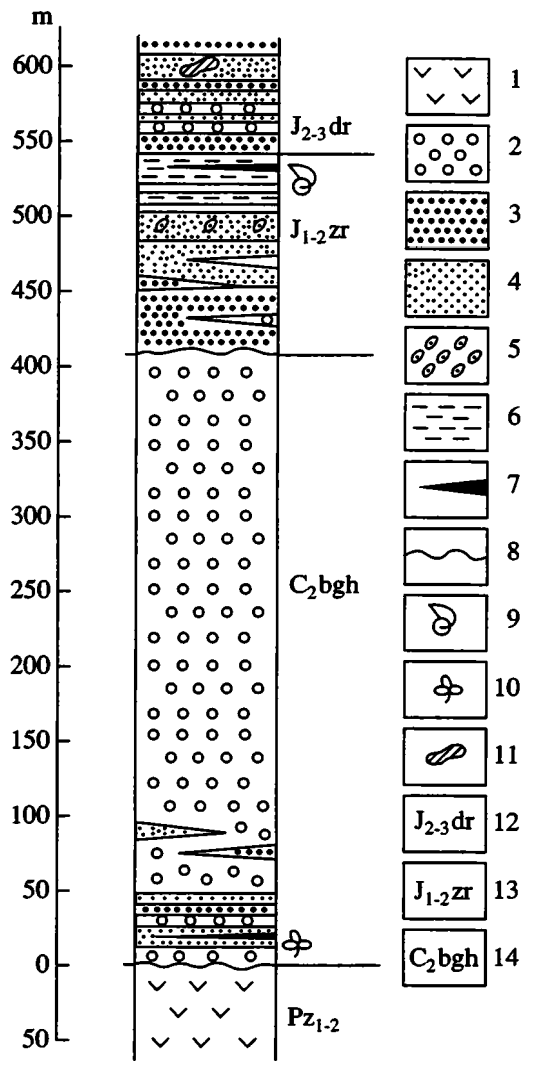

Fig. 2. Sequence of Paleozoic and Jurassic rocks on the northwestern slope of the Baga Khairhan Mt.

1 – Lower-Middle Paleozoic volcanic rocks, 2 – conglomerates, 3 – gravelstones, 4 – sandstones, 5 sandy ball-like concretions, 6 – claystones and shales, 7 – coal beds and lenses, 8 – unconformity, 9 – samples with molluscan fossils, 10 – samples with plant remains, 11 – fragments of dinosaurs bones, 12 – the Darbi Formation, 13 – the Zhargalant Formation, 14 – the Baga Khairhan Formation.

OLIGOCENE DEPOSITS

The Paleogene red deposits mainly occur in the western and southwestern areas of study, where they are widespread between the Gurvan Range (the outer part of Mongol Altai) and the southern Sutug Ula Range (the axial part of Mongol Altai) (Fig. 1). The former is composed of Mesozoic deposits, and the latter comprises Paleozoic rocks. The main Paleogene outcrops are located west and south of the Gurvan Eren Range. Here, the Paleogene strata unconformably overlie paleontologically dated Cretaceous (Aptian–Albian) deposits (the Dzereg Formation). In addition to an evident stratigraphic hiatus between the units, an angular unconformity is observed locally. For example, southwest of the Guen Khuduk Well, the angular unconformity is up to 10° - 12°. The dip angles of the Paleogene layers vary from 2° to 7° to the southwest (3° - 5° on average), whereas those of the Dzereg Formation beds are 10° - 30°.

The most complete and representative Paleogene section was observed 1.5 km south of the Guen Khuduk Well. In this sequence, the underlying grey Aptian–Albian deposits are predominantly composed of sandstones with claystone intercalations that contain the

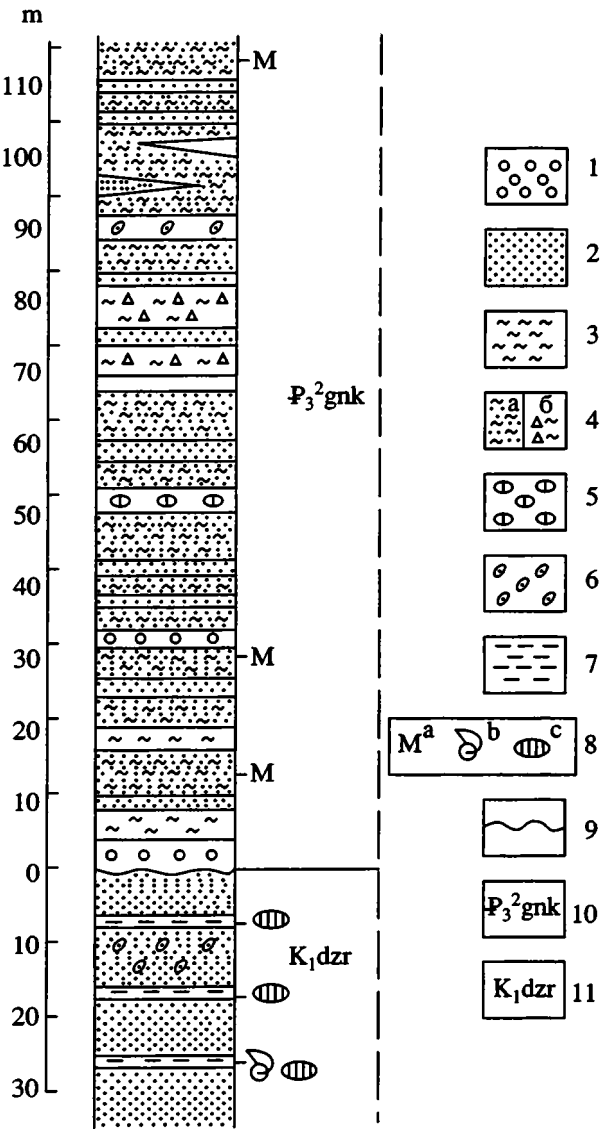

Fig. 3. Sequence of Paleogene and Lower Cretaceous rocks southwest of the Guen Khuduk Well.

1 – conglomerates; 2 – sandstones; 3 – clays; 4 – sandy (a) and debris (b) clays; 5 – carbonate concretions; 6 – sandy concretions; 7 – claystones; 8 – organic remains: a – mammalian, b – mollusks, c – ostracods; 9 – unconformity; 10 – the Guen Khuduk Formation; 11 – the Dzereg Formation.

remains of mollusks and ostracods (Fig. 3). The Paleogene sequence that I consider as the separate Guen Khuduk Formation incorporates the following strata (from the base upward):

(1) Brownish-grey weakly cemented conglomerates. The thickness is 3 - 4 m.

(2) Red-brown clays with intercalations of grey sandstones, 0.3 m thick. The clays yielded fragments of bones and teeth of mammalian *Tsagonomis altaicus* Matthew et Granger and others. The thickness is 15 m.

(3) Red-brown sandy clays with interbeds of small-pebble conglomerates and grey-brown sandstones, up to 1 m in thickness. In the clayey layers, I have found bone fragments, jaws, and individual teeth of mammalian *Cyclomylus lohensis* Matthews et Granger, *Desmatolagus* sp., and *Karakorumys* sp. The thickness is 20 m.

(4) Red-brown and reddish-brown clays with an admixture of debris and sand. Isolated interbeds, up to 0.5 m in thickness, composed of dense pinkish-brown sandstones and carbonate concretions occur.

The clay layers incorporate poorly preserved fragments of mammalian bones. The thickness is 30 m.

(5) An alternation of brownish-red and reddish-brown sandy clays with gravel grains and interlayers, up to 1 m in thickness, of weakly cemented sandstones and carbonate concretions. Single lenslike beds, up to 0.5 m in thickness, of small-pebbled greyish-brown conglomerate are observed. Bone fragments and teeth of mammalian *Tsagonomys altaicus* Matthew et Granger were sampled in the clayey beds. The thickness is 30 m.

(6) Dark red-brown and brownish-red clays with intercalations and lenses of gravelstone and grey small-pebbled conglomerate.

The clayey layers yielded remains of mammalian *Tsagonomys altaicus* Mathew et Granger and *Cyclomylus lohensis* Mathews et Granger. The thickness is 10 m. Quaternary grey deposits, 1 - 2 m thick, unconformably overlie this bed.

The total visible thickness of the Guen Khuduk Formation is 100 - 110 m.

Six km southeast of the Guen Khuduk Well, I have observed deposits similar in composition and color. Here, the upper layers of the Aptian–Albian section dip southwest at an angle of 10° - 12°. They are represented by ostracod-bearing claystones and grey sandstones with coprolites and ostracods. The thickness of individual layers varies from 0.5 to 3 - 5 m. Between these rocks and unconformably overlying the Paleogene red deposits a member of brownish-red and brownish-yellow clays is observed that resembles the Lower Cretaceous weathering crust. The possibility can not be excluded, however, that these layers belong to the uppermost Dzereg Formation. The thickness of this unit is about 4 m.

The sequence of red deposits of the Guen Khuduk Formation here incorporate the following strata (from the base upward):

(1) Basal conglomerates composed of various pebbles of low-middle roundness class, which are cemented by a sandy gravel matrix. The rocks include lenses of grey sandstones with varying grain size, gravelstones, and light-grey calcareous sandstones with an admixture of rare pebbles. The thickness is 10 m.

(2) Brown sandstones and gravelstones. The thickness is 3 m.

(3) Red-brown and bright red dense clays, locally sandy, with intercalations and lenses of sandstones with

varying grain sizes and small-pebble conglomerates. In the middle part of the bed, bone remains of mammalian *Tsegonomys altaicus* Matthews et Granger were sampled. The thickness is 30 m. The rocks dip southward at an angle of 2° - 7° and tend to flatten horizontally toward the southwest. The higher layers were not observed. The total visible thickness of this section exceeds 40 m.

Small outcrops of similar rocks were also observed south of the Baga Khairhan Mt. and in some other localities in this area.

I suggest that the section located 1.5 km south-southwest of the Guen Khuduk Well is the stratotype, because it is both the thickest and the most completely characterized paleontologically. The middle Oligocene age of this formation is supported by mammalian remains.

These deposits are very similar to those of the Shand-Gol Formation described by Devyatkin (1981) and Dashzeveg (1970) in the Valley of Lakes, which are dated as the middle Oligocene. After studying my collection, Dashzeveg noted that the fauna sampled in these deposits is identical to that from the Shand-Gol and Tata-Gol formations and the fauna from the Grand Canyon locality in North America, and that it is definitely the middle Oligocene in age.

CONCLUSION

Thus, new data on the Paleogene deposits provide a sufficient basis for assigning them not only to the Paleogene, as was assumed earlier (Devyatkin *et al.*, 1975), but also to a narrower time interval, i.e., the middle Oligocene. These rocks are widespread in both the western and central areas of Mongolia as well as its southern and southeastern areas (Dashzeveg, 1970; Devyatkin, 1981).

The first findings of Carboniferous flora and Oligocene fauna in the Darbi locality allows us to propose it as the reference section for studying not only the Mesozoic deposits, as was agreed earlier, but the Paleozoic and Cenozoic continental sequences of Mongolia and adjacent Asian regions as well.

Reviewer M.A. Akhmet'ev

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In Commemoration of Vladimir Vladimirovich Tikhomirov, Corresponding Member of the Russian Academy of Sciences (1915 - 1994)

Vladimir Vladimirovich Tikhomirov, an outstanding scientist with wide geological interests and an eminent researcher noted in the history and methodology of the earth sciences, died on January 13, 1994.

He was born in Petrograd on October 25, 1915, into the family of a teacher of physical chemistry, who later became a professor and an Honored Scientist of Azerbaijan. Graduating from the Baku Industrial Institute, Geological Prospecting Faculty, in 1938, Tikhomirov became a geological engineer in the Azerbaijan Geological Department and then the head of geological prospecting and surveying parties. In 1940, Tikhomirov joined the Communist Party, remaining a member for his entire life. Staying with the Azerbaijan Geological Department, he estimated commercial resources and assessed some mineral deposits and also conducted a geological survey of separate areas of the Transcaucasian region for a short period of time. From the beginning of World War II, Tikhomirov was engaged in providing heavy industry with raw material.

In May 1942, he joined the Field Army as a volunteer. After training in the military course at the Lenin Military Political Academy, he served in the 13th Air Force Army of the Leningrad Front, participating in air flights as the assistant commander of a detached air squadron and a flight navigator. In April 1944, Senior Lieutenant Tikhomirov was seriously wounded, resulting in his complete loss of sight; after a long treatment period, he was demobilized in January 1945. Tikhomirov entered a new phase of his life with fortitude, and he was able to master a sufficient strength of will to continue his scientific work. After leaving the hospital, he joined a post-graduate course at the Moscow Geological Prospecting Institute, and in 1949, he defended a dissertation devoted to the geology of the Lesser Caucasus. The Scientific Council of the institute awarded Tikhomirov with the Candidate and Doctorate degrees in Geology for this study. The work summed up his prewar investigations, which were supplemented with new data on the Caucasus geology obtained during the postwar period. A method used in the thesis of compiling stage-by-stage paleogeographic maps for the study of the geotectonics of the region aroused the interest of geologists concerned with facies analysis, paleogeography, and tectonics. In 1953, The Moscow Society of Naturalists (MSN) awarded the first prize to Tikhomirov's monograph *Lesser Caucasus in the Late Cretaceous* (1950).

In 1949, N.S. Shatskii invited Tikhomirov to the Institute of Geological Sciences of the USSR Academy of Sciences to organize and conduct planned investigations on the history of geological sciences. In 1951, the Cabinet of the History of Geological Sciences was founded with Tikhomirov as the head, which was later reorganized into the Department, and then the Laboratory of the History of Geology in the newly formed Geological Institute (GIN) of the USSR Academy of Sciences. In 1955, the professor degree in the history of geological sciences was conferred to him. In 1956, Tikhomirov, along with V.E. Khain, issued *A Brief Essay on the History of Geology*, which was widely used as a textbook at the geological faculties of different institutes and was translated into Chinese in 1959. At the beginning of the 1960s, Tikhomirov's monograph *Geology in Russia in the First Part of the 19th Century* was published in two volumes (1960 and 1963), containing a comprehensive analysis of this science for a poorly studied period. Along with actual data, the book

comprises theoretical propositions on regularities in the development of the science, and for the first time, the evolution of scientific and philosophical views of Russian geologists was considered. In 1965, the MSN awarded first prize to this work.

Tikhomirov was the initial creator and chief-editor of the well-known series *Essays on the History of Geological Knowledge*, in the framework of which 28 books have been published since 1953. Issue 20 (1979) represents Tikhomirov's monograph *Geology in the Academy of Sciences (from Lomonosov to Karpinskii)*.

In 1963, Tikhomirov was elected a Corresponding Member, and in 1966, a Member of the International Academy of Science History (Paris) for his outstanding achievements in developing studies on the history of geology. He was the only geologist among its members.

On academician V.A. Obruchev's initiative, the Commission on Geological Study of the USSR (CGS) was founded in 1955 with the immense objective of revealing and gathering materials related to all geological investigations ever conducted in this country. At the very beginning, Shatskii (Director of the Geological Institute) was the Chairperson of CGS, Tikhomirov, its Vice president, and Yu.Ya. Solov'ev, its Scientific Secretary. Shortly thereafter, on the basis of his talent for organization, Tikhomirov was appointed the Chairperson of CGS and editor-in-chief of the 52-volume issue *Geological Study of the USSR*. To prepare this issue, the CGS collaborators supervised in the 1960s the work on data collection, processing, and systematization of all geological investigations carried out in the territory of the FSU, as well as the summarization of the initial data gathered. As a result, we now have the reference and information issue *Geological Study of the USSR*, unique in its material scope and coverage of the area. It would be hard to find its match in any other field of knowledge or in any other country. Currently, 150 books of this issue have been published, which are used not only by geologists, but economists, geographers, historians of geology, bibliographers, and other specialists as well. The role of Tikhomirov in the activities of the CGS is hard to overestimate. It should be pointed out that the interest of geologists in the history and methodology of geological sciences at the end of the 1960s and the beginning of the 1970s was aroused by the activity of Tikhomirov and his laboratory staff. This group was always a competent informational center, uniting the activity of geology historians from 36 countries and 5 continents. The authors of the jubilee monographs *History of the Geological Institute of the USSR Academy of Sciences: Evolution of the Institute, Its Scientific Schools, and Bibliography* (1980), headed by Tikhomirov, showed that a retrospective analysis of the development of fundamental trends in geology enables us to evaluate the present state of the science and predict its future. Due to Tikhomirov's efforts, a new branch of the earth sciences, the history of geological sciences, developed. Equally with other traditional geological sections, the Department of the History of Geology has been included in the programs of International Geological Congresses since the 23rd Session (Prague, 1968).

In the 1980s, the Presidium of the USSR Academy of Sciences extended the sphere of responsibilities of the CGS and suggested that the Commission should coordinate the activity of individual scientists and groups of scientists dealing with the history and methodology of geological sciences, organize work on the study of the history of geological investigations in the USSR, methodologically supervise the compilation of reference-bibliographical issues, and participate in the organization of international and domestic conferences on the history and methodology of geological investigations. In 1983, the CGS was reorganized into the Commission of the Academy of Sciences on the History of Geological Knowledge and Geological Study of the USSR, and Tikhomirov, who was elected a Corresponding Member of the Academy of Sciences in 1981, was again appointed to be its chairperson.

New facets of Tikhomirov's talent for organization manifested themselves in the activities of the International Commission on the History of Geological Sciences (ICHGS), founded in 1967. During the years 1967 - 1975, Tikhomirov was twice elected its president, and since 1984 he was the vice-president. Tikhomirov was the initiator of some major international and domestic symposia on the history and methodology of geological sciences and the chairperson of the section "History of Geology" at the 27th Session of the International Geological Congress (Moscow, 1984) and the 11th ICHGS Symposium. Owing to his inexhaustible efficiency, erudition, and vitality, Tikhomirov became a well-known organizer and leader of investigations into the history of geological sciences, as well as a prominent geologist. He was a permanent consulting editor of the mining and geological section of the *Greater Soviet Encyclopedia* and the reference issue *The Mining Encyclopedia*. In 1976, the Society of Geological Sciences in Berlin elected Tikhomirov an Honored Member.

Giving much to scientific and organizational activity in the field of the history of geology, Tikhomirov was also concerned with modern problems of the earth sciences. At the end of the 1950s and the beginning of the 1960s, he carried out investigations on theoretical problems of geotectonics, and at the beginning of the 1970s, he developed the original views of V.I. Vernadskii and L.S. Berg about the evolution of microorganisms in the Early Archean.

Tikhomirov analyzed in detail the peculiarities of the evolution of ideas and methods in the history of geology and presented its periodization in a new manner. He often emphasized that periodization was not an artificial method of research but an objective reality, resulting from the nonuniform development of the science.

In December 1987, according to government resolution, the V.I. Vernadskii State Geological Museum was founded in the central part of Moscow; in March 1991, the Presidium of the Academy of Sciences considered it expedient to transfer the Department of the History of Geology from the Geological Institute to this unique museum in order to further develop the study of the history of this science. Simultaneously, the important task of using the valuable stock of historical and scientific information collected in museums, expositions, and exhibitions was proposed. As in the past, Tikhomirov made long-term plans to collect material for the theme "The Culture of Stone in Ancient Eurasia." In 1992, Tikhomirov became a member of the Russian Academy of Natural Sciences.

About 400 works published by Tikhomirov, either alone or with co-authors, testify to the striking productivity of his scientific activity. Many works have been translated into foreign languages and published in 18 countries. Contemporaries of Tikhomirov and the following researchers will inevitably make use of the scientific heritage left by Tikhomirov.

The outstanding scientific service and successes achieved by Tikhomirov were recorded with orders: Order of the Patriotic War of the 1st and 2nd Range, Order of the Red Star, the Badge of Honor, Order of the Red Banner of Labor, Order of the October Revolution, and many medals.

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